Architecture, 3D geometry and tectonic evolution of the Carpathians foreland basin

Tarapoanca, M.

2004

VRIJE UNIVERSITEIT

ARCHITECTURE, 3D GEOMETRY AND TECTONIC EVOLUTION OF THE CARPATHIANS FORELAND BASIN

ACADEMISCH PROEFSCHRIFT

ter verkrijging van de graad van doctor aan
de Vrije Universiteit Amsterdam,
op gezag van de rector magnificus
prof.dr. T. Sminia,
in het openbaar te verdedigen
ten overstaan van de promotiecommissie
van de faculteit der Aard- en Levenswetenschappen
op dinsdag 13 april 2004 om 13.45 uur
in het auditorium van de universiteit,
De Boelelaan 1105

door

Mihai Tărăpoancă

geboren te Boekarest, Roemenië

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The research reported in this Thesis has been carried out at:

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ISES

The research was funded by the Netherlands Centre for Integrated Solid Earth Sciences (ISES) at Vrije Universiteit.

This is publication no. 2004.01.03 of the Netherlands Research School of Sedimentary Geology (NSG).

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ISBN 90-9017847-3

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ACKNOWLEDGEMENTS

This Thesis has been accomplished thanks to the support and help of many people that I am indebted to. This section is dedicated to them and I will mention and thank those who have contributed in various ways to either scientific work or personal support.

I begin by thanking Prof. Sierd Cloetingh, who kindly acquired me in the Tectonics Department, where I have found an excellent atmosphere, both in terms of professional expertise and human quality. During all my research stages in Vrije Universiteit, I have received permanent encouragement from him and also, benefited a lot from his suggestions and comments that improved substantially the quality of the Thesis. Last, but not least, I thank Prof. Cloetingh very much for his involvement in my professional future.

I am also greatly indebted to my promoter from Bucharest University, Prof. Corneliu Dinu. From the times when I was a student, he has given permanent support to me, and I have always found his door open. I thank him for accepting me to begin the PhD under his promotion, and, afterwards, for his recommendations that allowed me to continue the work for the Thesis at Vrije Universiteit.

Many thanks go to Dr. Giovanni Bertotti, who was directly supervising my research work during the visits at Vrije Universiteit. I have learned a lot of very useful things from him, from how to address a problem, to writing comprehensibly a paper, to better organising a scientific approach and many others. He put considerable effort in guiding me through all the steps towards the end of Thesis, and it is my great pleasure to thank him for the important contributions that he has had to my professional achievements.

Dr. Liviu Mațenco is the person that I owe to many of the advancements towards the present Thesis. Since the times when "you're forbidden to come in my office unless you have at least one page written", he has strongly encouraged and "pushed" me in so many ways and so many times, that this section would be too short to mention all. I thank him a lot for these, and especially for his friendship. He and Ioana have been always ready to help me to solve any problem I had.

The Thesis has got important improvents due to the valuable suggestions and comments made by the reading committee: Dr. O. Dicea, Prof. H. Doust, Dr. F. Roure, Dr. R. Stephenson, Dr. M. Ştefănescu and Prof. P. Ziegler. I thank to all of them for their effort and time spent for reading and correcting the manuscript. In addition, the major contribution that Dr. Oprea Dicea had upon my professional background during the period worked in Prospecțiuni S.A. is greatly acknowledged. I got very useful insights in the modeling principles and techniques from the course held by Dr. Randell Stephenson, for which I thank him very much. Also, I acknowledge the financial support received from the ISES project managed by Dr. Stephenson. I am very indebted to Dr François Roure, who invited me for a post-doctoral stage in Institut Français du Pétrole.

The periods spent in Amsterdam would have been difficult without many of the people that I have met here and with whom I have enjoyed a real pleasant time. I start by remembering with great pleasure those nice evenings with Dimitrios. I found in my friend many common points, from eating extremely hot to the passion for history, not only the earth one, but also that of the mankind. Jeroen introduced me in the beauty of Amsterdam (honni soit qui mal y pense) and his friendship is much appreciated. Also, I am indebted to him and to Kareen for translating the summary of the manuscript. Tristan, Violaine and the little girls have provided me a warm environment, each time when we could have some time together. I appreciate very much the contribution that Daniel had upon the modeling part of the Thesis and I remember with great pleasure the time spent together. Also, I have very much enjoyed the company of Jorge, Stéphanie, Gabor B., Laszlo, Gabor W, Sevgi, Geza. Fred, Ernst, Anco, Reini, Bernd, Ingrid, Marlies, Dick, Christophe were ready to help me anytime I needed or to answer to any question I had, and I thank to them for that. I acknowledge the Alwine's help in some administrative tasks.

I thank a lot to my friends from Bucharest University that "hosted" me for pretty long time, for discussions (not only about geology), providing many good solutions to various problems (as those related to computers) or sharing ideas, and mainly, for their friendship. These acknowledgements warmly go to Coco, Costica, Cristi, Daniel, Diana, Gigi, Iuliana, Lili, Marius S., Marius T., Mihai, Pongo, Sebi, Victor and Vlad. In addition, Victor is thanked for the support received at the S. Mueller conference in Cheile Buții.

I have got a lot of experience and I have learned many things during the seven years worked in Prospecțiuni S.A., where I had a nice work environment provided by my former colleagues. I keep good memories from that period and I would like to thank to Alex, Bebe, Cori, Daiana, Dan, Emil, Gabi, Geta, Jean, Mihai, Nelu, Rodica, Romeo, Tavi, Victor and all the others. I greatly appreciate the friendship and help of Traian and Anamaria, as well as all the geological talks that we have had since we were students. Also, I thank for geological (and not only) discussions and their warm presence to Adriana, Aida, Dorina, Ion and Justin from Petrom S.A.

I would like also to express my reconnaissance to my first professor in geology, who transmitted me the passion for the earth history and, somehow, outlined my professional future: Prof. Nicolae Ungureanu from Petroşani University.

Of course, the most important support has been provided by my wife Laura and my parents. I do not have enough words to describe this, as well as to express my gratitude. This book is written also for them and for my child.

SAMENVATTING

Dit proefschrift is gericht op de mechanismen die het patroon van verticale bewegingen in voorlandbekkens bepalen en op invloed van laterale veranderingen in de mechanische eigenschappen van de voorlandplaat. Veranderingen in ruimte en tijd van de daling en structurele stijl van voorland bekkens zijn ook verbonden met de deformatie geschiedenis van de nabije Karpatenboog, en de studie van deze veranderingen draagt ook bij aan een beter begrip van de syn- en post- compressieve fasen van orogene systemen. Implicaties voor de processen die actief zijn in dit type tektonische systemen worden afgeleid van de Roemeense voorland/voordiep bekkens omdat zij getuigen van een snelle opeenvolging van verschillende mechanismen, waaronder flexuur van de voorlandplaat gedreven door orogene belasting op die plaat en extensie tot lithosferische plooiing, in het kader van de algemene convergentie. Op bekkenschaal wordt het relatieve belang van elk van deze mechanismen grotendeels gecontroleerd door het lithosferische geheugen van de voorlandplaat. Ofwel, de gevolgen van eerdere tectonische processen kunnen blijvend zijn, en aldus de lokalisatie en mate van daling/deformatie bepalen.

De Tertiaire tektonische evolutie van de Zuid en Oost Karpatische voorland/voordiep bekkens wordt afzonderlijk behandeld vanwege de afwijkende tektoniek en patronen van verticale bewegingen in deze gebieden. De in dit proefschrift gepresenteerde onderzoeksresultaten zijn grotendeels gebaseerd op de interpretatie van seismische reflectie lijnen die verkregen en bewerkt zijn door de olieindustrie. De interpretatie van deze seismische lijnen heeft ons in staat gesteld om de architectuur van het Karpatische voorland/voordiep bekken te definiëren de geometrische veranderingen in het bekken te verduidelijken. De structurele kenmerken van de voorlandplaat, alsmede zijn deformatie geschiedenis, kunnen worden gedetailleerd en verder geïntegreerd op regionale schaal. Na de evolutie van het voorland bekken ontrafeld te hebben, zijn de implicaties voor de tektonische processen die daaraan ten grondslag liggen afgeleid en vervolgens getest met behulp van kwantitatieve modellering. Modellering van de dalingsgeschiedenis van het Karpatische voorland/voordiep is toegespitst op de Focşani depressie. Deze depressie is onderwerp van menig debat betreffende zijn extreme diepte (~13 km) en genese, evenals zijn externe positie ten opzichte van het orogeen.

Hoofdstuk 1 definieert de doelstellingen van dit proefschrift en de belangrijkste vraagstukken van de bestudeerde regio. Daarna volgt een korte beschrijving van de inhoud van elk hoofdsstuk.

Hoofdstuk 2 presenteert een synthese van de structurele evolutie van het Karpaten gebergte en zijn voorland/ voordiep. De tectonische evolutie van het zuidelijke en oostelijk deel van de Karpaten wordt in een tijdskader geplaatst, met de nadruk op de Neogene deformatie. In een volgende stap wordt het Karpatische voorland beschreven dat bestaat uit vier lithosfeer blokken, te weten het Oost-Europese, het Scytische, en het Moesische platform en het Noord Dobrogea gebergte. Verschillende hypotheses betreffende de algemene evolutie van de Karpaten en de daling gelieerd aan de buigzone worden kort besproken. In het algemeen hebben deze hypotheses gemeen dat de Neogene evolutie van het Karpaten gebergte hoofdzakelijk wordt bepaald door de “roll-back” en ontkoppeling van oceanisch of continentaal mantelmateriaal dat gekoppeld was aan de voorlandplaat. Het wordt algemeen aangenomen dat dit proces ook verantwoordelijk is voor het vulkanisme aan de interne zijde van het Karpaten gebergte, de anomale daling van de Focşani depressie en de seismiciteit in het Vrancea gebied. Maar het patroon van verticale bewegingen in ruimte en tijd suggereert dat andere mechanismen (zoals lithosferische plooiing) wellicht ook een rol van betekenis spelen.

Hoofdstuk 3 refereert aan de evolutie van het Zuid Karpatische voordiep en voorland, oftewel het westelijke gedeelte van Moesië (ten westen van de Intra-Moesië breuk). De zuidelijke marge van Moesië evolueerde sinds het Eoceen als het voordiep van de Balkan terwijl de noordelijke marge onder invloed van extensie was, in ieder geval tijdens het Vroeg Mioceen. Deze noordelijke marge ontwikkelde zich pas tot het voordiep van de Zuid Karpaten tijdens het Laat Mioceen (Sarmatien) tijdens de plaatsing van het Zuid Karpatische dekblad. Tussen de twee dalingsfasen geassocieerd met de flexuur van Moesië was grootschalige erosie het dominante proces. Een netwerk van valleien en canyons met een breedte van soms tientallen kilometers en enkele kilometers diep werd gevormd als antwoord op de opheffing van Moesië die het resultaat is van de gecombineerde van het Balkangebergte en/ of een flexurale “forebulge” gevolgd door Vroeg Miocene opheffing van de rifschouder. Hoewel de opschuiving van het Sub-Karpatische dekblad plaatsvond tijdens het Laat Mioceen, ging de daling in het voorland nog lange tijd door. Deze post-erosie daling wordt beschouwd als het gevolg van de flexuur van het voorland gevolgd door plooiing van de lithosfeer.

Hoofdstuk vier behandelt het voorland van de Oost Karpaten, met de nadruk op de gedetailleerde structurele, en dalingsgeschiedenis van de Focşani depressie. Door zijn ligging op de kruising van het Oost Europese en het Scytische platform, het Noord Dobrogrea gebergte (promontory) en oostelijk Moesië, toont dit bekken de grootste daling van het gehele Karpatische voordiep. De tectonische evolutie en verticale bewegingen van het voorland ten noorden van de Trotus breuk (Oost-Europese/Scytische platform) is duidelijk verschillend van die van de buigzone van de Karpaten (Moesië): waar de structurele stijl van het voordiep van het Oost Europese/Scytische platform de mechanische koppeling tussen het orogeen en het voorland reflecteert (compressie overgebracht naar het voorland), was het structurele patroon in Moesië extensioneel (NW-ZO georiënteerd) gedurende het Midden Mioceen en vervolgens “strike-slip”. Bovendien, de daling in het Oost-Europese/ Scytische platform eindigde na de laatste fase van gebergte vorming. Daartegenover staat dat de daling van Moesië doorging na de voornaamste fase van gebergte vorming en zelfs toenam vanaf het Plioceen. De flexuur van de voorlandplaat als resultaat van de orogene belasting lijkt de daling van het Oost-Europese/ Scytische platform volledig te kunnen verklaren, en draagt bij aan de daling van Moesië. Extensie voorafgaand aan de flexuur is het voornaamste proces dat heeft geleid tot de daling van Moesië. De ruimtelijke distributie van de jongste, Plio-Kwartaire, verticale bewegingen duidt erop dat deze gerelateerd zijn aan plooiing van de lithosfeer.

Hoofdstuk 5 bevat een kwantitatieve analyse van de daling van het voordiep van de Karpaten gebaseerd op numeriek modelleren van de buiging (flexuur) en extensie van de lithosfeer. Die laatste numerieke methode houdt rekening met laterale variaties in de sterkte van de lithosfeer en met de werkelijke geometrie van de topografische belasting. Hoewel van ondergeschikt belang**,** droeg de extensie in de zuidoostelijke Karpaten, samen met het effect van lineamenten die actief zijn op de schaal van de gehele korst, bij aan het significant verzwakken van de lithosphere van oostelijk Moesië. De aanwezigheid van deze zwakke lithosfeer speelt een cruciale rol in de lokalisatie van de belangrijkste dalingscentra. Hier wordt aangetoond dat de globale vorm van het voordiepbekken en het merendeel van zijn diepte kunnen worden verklaard door de opeenvolging van twee mechanismen, te weten extensie en flexuur. Het resultaat suggereert ook dat lithosfeer plooiing aan het einde van de convergentie een valide mechanisme is voor de laatste dalingsfase, die voornamelijk is geregistreerd in gebieden met een zwakke lithosfeer.

In hoofdstuk 6 worden de conclusies uit de hoofdstukken 3 en 4 geïntegreerd in een regionaal kader voor verschillende perioden. Het blijkt dat, terwijl de interne Karpatische eenheden rond Moesië roteren en richting het noordoosten en oosten bewegen, de gehele noordelijke Moesische marge meer en meer aan extensie onderhevig was. Deze extensie migreerde van west (Vroeg Mioceen) naar oost (midden Mioceen) in combinatie met een rotatie rechtsom, gelijk aan die van de Karpatische eenheden. Verschillen in tectonische evolutie en verticale bewegingen tussen de voorlanden van Oost en Zuid Karpaten (Oost-Europese/ Scytische platform en Moesië respectievelijk) worden klaarblijkelijk gecontroleerd door de rheologische eigenschappen van de lithosfeer van het voorland (sterk versus zwak), evenals door de richting van de convergentie/ collisie (frontale versus diagonale collisie).

SUMMARY

This PhD Thesis focuses on mechanisms controlling the pattern of vertical movements in foreland basins and on the influence of lateral changes of foreland plate mechanical properties on the subsidence and tectonic histories. Temporal and spatial variations in subsidence and structural style of foreland basins are also genetically related to the deformation history in the neighbouring fold-and-thrust belt, and studying these variations contributes towards a better understanding of the syn- and post-collisional stages of orogenic systems. Inferences on the processes acting in this type of tectonic setting are derived from the Romanian Carpathians foreland/foredeep basins since they provide evidences for a rapid succession of different mechanisms, including flexure of the foreland plate driven by orogenic loading and extension to lithospheric buckling, in the framework of the overall convergence. At a basin scale, the relative importance of each of these mechanisms is largely controlled by the lithospheric memory of the foreland plate. That is, the effects of a tectonic process can be inherited, thus controlling the localization and amount of subsidence/deformation during the subsequent stages.

The Tertiary tectonic evolution of the foreland/foredeep basins of the South- and East-Carpathians are discussed separately because they are characterized by significant differences in terms of tectonics and patterns of vertical movements. Research results presented in this Thesis are largely based on the interpretation of reflection seismic lines acquired and processed for the petroleum industry. Interpretation of these seismic surveys has allowed us to define the architecture and highlight changes in geometry of the Carpathians foreland/foredeep basin. The structural style of the foreland plate, as well as its deformation history, can be detailed and further integrated at a regional scale. Having constrained the evolution of the foreland basin, inferences on controlling tectonic mechanisms have been derived and subsequently tested by means of quantitative modeling. Subsidence modeling of the Carpathians foreland/foredeep is focused on the Focşani Depression, which has been the topic of many debates concerning its extreme depth (~13 km) and genesis, as well as its external position relative to the orogen.

Chapter 1 defines the objectives of this Thesis and the main problems of the studied region, giving also a brief presentation of the contents of each chapter. Chapter 2 presents a synthesis of the structural evolution of the Carpathians orogen and its foredeep/foreland. The tectonic evolution of the South- and East-Carpathians sectors is placed within a temporal framework, emphasizing Neogene deformations. In a next step, the Carpathians foreland is described, which consists of four lithospheric blocks, namely the East-European, Scythian and Moesian platforms and the North Dobrogea orogen. Various hypotheses concerning the general evolution of the Carpathians and the subsidence associated with the Bend Zone are briefly discussed. Overall, these converge on the concept that the Neogene tectonic evolution of the Carpathians belt was essentially controlled by the roll-back and detachment of an oceanic or continental mantle slab that was attached to the foreland plate. Generally, it is assumed that this process is also responsible for the volcanism in the inner part of the Carpathians belt, the anomalous subsidence of the Focşani Depression and the seismicity of the Vrancea Zone. However, the spatial and temporal pattern of vertical movements suggests that other mechanisms (as lithospheric buckling) are likely to play an important role as well.

Chapter 3 refers to the evolution of the South-Carpathians foredeep and foreland, i.e. the western part of Moesia (to the west of the Intramoesian fault). The southern margin of Moesia evolved since the Eocene as the foredeep of the Balkans while its northern margin was affected by extension, at least during the Early Miocene, only developing into the foredeep of

the South-Carpathians in the Late Miocene (Sarmatian), when the Subcarpathian nappe was emplaced. Between the two subsidence stages associated with the flexure of Moesia, the dominant process was large-scale erosion. A network of valleys and canyons up to tens of kilometres-wide and a few kilometres-deep was formed in response to uplift of Moesia resulting from the spatial combination of the Balkans thrusting-induced arch and/or flexural fore-bulge followed by the Early Miocene rift shoulder. Although thrusting of the Subcarpathian nappe is Late Miocene in age, subsidence of the foreland continued for a long time afterwards. This post-erosional subsidence is seen as resulting from a foreland flexure followed by lithospheric buckling.

Chapter 4 deals with the East-Carpathians foreland, paying particular attention to the detailed structure and subsidence history of the Focşani Depression. Lying at the junction between the East-European/Scytian platform, North Dobrogea orogen (promontory) and eastern Moesia, this basin records the largest subsidence along the entire Carpathians foredeep. There is an obvious difference in terms of tectonic evolution and vertical movement between the forelands to the north of the Trotuş fault (East-European/Scythian platform) and of the Bend Carpathians (Moesia): whereas on the East-European/Scythian platform the structural style of the foredeep reflects mechanical coupling between the orogenic wedge and the foreland (compression transmitted to the foreland), the structural pattern in Moesia is extensional (NE-SW oriented) during the Middle Miocene and strike-slip afterwards. Also, the subsidence of the East-European/Scythian platform ceased after the last collisional event. By contrast, the subsidence of Moesia continued after the major collisional event and even increased starting in the Pliocene. The flexure of the foreland plate produced by orogenic loading appears to fully explain the subsidence in the East-European/Scythian platform, and partly contributes to the subsidence of Moesia. In the latter, extension prior to flexure is the main process involved in the initial subsidence stage. Based on the spatial pattern of the Pliocene-Quaternary vertical movements, the last subsidence stage appears to be related to lithospheric buckling.

Chapter 5 comprises a quantitative analysis of the Carpathians foredeep subsidence, based on extensional and planform flexural numerical modeling. The latter numerical approach takes into account lateral variations in lithospheric strength, as well as the actual geometry of the topographical load. Although minor, the extension in the SE Carpathians foreland, together with the effect of crustal-scale active lineaments, contributed to significant weakening of the eastern Moesian lithosphere. The presence of this weak lithospheric domain has a crucial impact on the localization of the main subsidence centres. It is demonstrated that the overall shape of the foredeep basin and most of its depth can be explained by a succession of two mechanisms, namely extension and flexure. The results also suggest that lithospheric buckling taking place in the aftermath of convergence can be a viable mechanism for the last subsidence stage, which is basically recorded in regions characterized by low lithospheric strength.

Chapter 6 integrates the conclusions drawn in chapters 3 and 4 within a regional framework for several time spans. It appears that, as the intra-Carpathians units rotated around Moesia and moved towards NE and E, the entire northern margin of Moesia was progressively affected by extension. This extension migrated from the west (Early Miocene) to the east (Middle Miocene) and its direction appears to have rotated clockwise, in tandem with the rotation of the Carpathians units. Differences in terms of tectonic evolution and vertical movements between the forelands of East- and South-Carpathians (East-European/Scythian platform and Moesia, respectively) were apparently controlled by the rheological properties of the foreland lithosphere (strong versus weak), as well as by the direction of convergence/collision (frontal versus oblique collision).

REZUMAT

Această Teză de Doctorat se concentrează asupra mecanismelor ce controlează mişcările verticale din bazinele de foreland, punînd de asemenea accent şi pe influenţa schimbărilor laterale de proprietăţi mecanice ale plăcii inferioare asupra istoriei subsidenţei şi tectonice. Variaţiile temporale şi spaţiale de subsidenţă şi stil structural ce caracterizează bazinele de foreland sînt legate genetic de istoria deformărilor din lanţul cutat învecinat. În consecinţă, studierea acestor variaţii reprezintă un pas important spre o înţelegere mai bună a perioadelor sin şi postcolizionale din sistemele orogenice. Informaţii despre procesele ce acţionează în acest tip de cadru tectonic se bazează pe studiul bazinelor de foreland ale Carpaţilor Româneşti deoarece aici se evidenţiaza o succesiune de mecanisme total diferite. În afară de flexura tipică a plăcii inferioare datorită încărcării orogenice, alte procese, cum ar fi extensia şi cutarea la scară litosferică, se pot observa în această regiune, toate fiind active într-un context general de convergenţă. La scară de bazin, ponderea fiecăruia dintre aceste procese depinde în principal de memoria litosferică a plăcii inferioare. Adică, efectele unui proces tectonic pot fi moştenite şi astfel să controleze localizarea şi magnitudinea subsidenţei/deformărilor din stadiile următoare. Foarte probabil, interacţiunea dintre diferite tipuri de procese tectonice implicate în arhitectura foreland-urilor devine cu atît mai importantă cu cît se referă la acele bazine asociate limitelor foarte curbate de plăci.

Intervalul de timp studiat este Terţiarul iar avanfosa Carpaţilor Meridionali este tratată separat de cea a Carpaţilor Orientali datorită diferenţelor semnificative legate de tectonică şi istorie a subsidenţei. Studiul se bazează în special pe interpretarea secţiunilor seismice de reflexie achiziţionate şi procesate pentru industria de petrol. Interpretarea acestor secţiuni permite identificarea cu acurateţe a arhitecturii şi a schimbărilor în geometria bazinului de foreland carpatic. Stilul structural al foreland-ului şi istoria deformării sînt detaliate şi apoi integrate într-un context regional. Avînd stabilită evoluţia foreland-ului, se determină mecanismele tectonice care sînt apoi testate prin modelare cantitativa. Modelarea cantitativa a subsidenţei pune accent pe Depresiunea Focşani care a prilejuit de-a lungul timpului cele mai multe dezbateri legate de adîncimea sa extremă (~13 km), geneză cît şi datorită poziţiei sale externe în raport cu orogenul.

Capitolul 1 cuprinde obiectivele Tezei, principalele probleme ale regiunii studiate precum şi o scurtă prezentare a conţinutului capitolelor lucrării.

În Capitolul 2 se face o trecere in revistă a cadrului structural al orogenului Carpatic şi avanfosei sale. Evoluţia tectonică a celor 2 sectoare ale Carpaţilor, Meridionali şi Orientali, este prezentată într-un cadru temporal, cu accent pe deformările Neogene. Urmează apoi prezentarea "foreland"-ului Carpaţilor, care este format din patru blocuri litosferice: Platformele Est Europeană, Scitică, Moesică la care se adaugă Orogenul Nord-Dobrogean. Sînt prezentate succinct opiniile despre evoluţia generală a Carpaţilor sau despre subsidenţa accentuată asociată Zonei de Curbură. În general, acestea converg spre ideea că elementul esenţial în evoluţia tectonică Neogenă a lanţului Carpatic este procesul de retragere şi detaşare a porţiunii de litosferă oceanică presupus ataşată de placa inferioara. În majoritatea opiniilor, acest proces este de asemenea responsabil şi de vulcanismul din zona internă, subsidenţa anormală din Depresiunea Focşani şi seismicitatea din Zona Vrancea. Cu toate acestea, distribuţia în timp şi spaţiu a mişcărilor verticale sugerează alte mecanisme, care vor fi discutate pe larg în cursul Tezei.

Capitolul 3 se referă la evoluţia avanfosei şi "foreland"-ului Carpaţilor Meridionali, adică a părţii vestice a Moesiei (la vest de falia Intramoesică). Marginea de sud a evoluat din Eocen ca avanfosă a Balcanilor în timp ce marginea nordica a suferit o extensie, cel puţin în

timpul Miocenului inferior, ca apoi să devină avanfosa Carpaţilor Meridionali începînd cu Miocenul superior (Sarmatian) cînd a fost amplasată pînza Subcarpatică. Între cele două stadii de subsidenţă asociate cu flexurarea Moesiei spre sud şi apoi spre nord, procesul dominant a fost eroziunea pe scară largă. O reţea de văi şi canioane care ajung la adîncimi şi lăţimi de ordinul kilometrilor şi, respectiv, zecilor de kilometri, s-a format datorită ridicării provocate de suprapunerea spaţiala a "fore-bulge"-ului Balcanilor şi ulterior, a umărului de rift Miocen inferior. Deşi încălecarea este Miocen superioară, subsidenţa a continuat mult timp după aceea. Această perioadă post-erozională este privită ca un rezultat al flexurii urmate de o cutare la scară litosferică.

Capitolul 4 tratează "foreland"-ul Carpaţilor Orientali, cu accent pe detalierea structurii şi istoriei subsidenţei în Depresiunea Focşani. Aflată la joncţiunea dintre Platforma Est Europeană/Scitică, Orogenul (Promontoriul) Nord-Dobrogean şi partea estică a Moesiei, acest basin înregistrează subsidenţa maximă din întreaga avanfosa Carpatică. Din punct de vedere al evoluţiei tectonice şi al mişcărilor verticale se constată o diferenţă netă între zona aflată în faţa Carpaţilor de la nord de falia Trotuşului (Platforma Est Europeană/Scitică) şi cea din faţa Carpaţilor de Curbură (Moesia). În timp ce în Platforma Est Europeană/Scitică stilul structural reflectă cuplarea din punct de vedere mecanic cu prisma orogenică iar subsidenţa încetează după momentul final de coliziune, în Moesia stilul structural este extensional (pe direcţie NE-SV) în Miocenul mediu urmat de cel de tip decroşare, iar subsidenţa continuă după momentul major de coliziune şi chiar se accentuează începînd din Pliocen. Flexura foreland-ului produsă de încărcarea orogenică pare să justifice în întregime subsidenţa în Platforma Est Europeană/Scitică şi parţial în Moesia. În aceasta din urmă, extensiunea anterioară flexurii reprezintă principalul proces implicat în primul stadiu de subsidenţă. Pe baza distribuţiilor mişcărilor verticale Pliocen-Cuaternare, ultimul stadiu de subsidenţă pare a fi legat de o cutare la scară litosferică.

Capitolul 5 cuprinde o analiză cantitativă a subsidenţei în avanfosa Carpaţilor, abordare realizată printr-o modelare numerică extensională urmată de una flexurală calculată în trei dimensiuni. Această ultimă modelare în 3D ţine seama de variaţiile laterale în rezistenţa litosferei precum şi de geometria reală a încărcării topografice. Deşi minoră, extensia din foreland-ul Carpaţilor de Curbură contribuie împreună cu aliniamentele structurale crustale active la slăbirea semnificatică a litosferei părţii de est a Moesiei. Prezenţa acestui domeniu litosferic slăbit are o importanţă crucială în localizarea bazinului principal la exteriorul centurii muntoase. Se demonstrează că forma generală şi cea mai mare parte a adîncimii bazinului de avanfosă se explică prin suprapunerea a două mecanisme: extensie şi flexură. De asemenea, rezultatele sugerează că ultimul stadiu de subsidenţă care este înregistrat în general în zonele cu rezistenţă redusă, poate fi legat în mod viabil de un proces de cutare la scară litosferică.

In Capitolul 6 se integrează concluziile din capitolele 3 şi 4 într-un cadru regional, prezentarea făcîndu-se pe cîteva intervale de timp. Se constată că pe măsură ce unităţile intra-Carpatice s-au rotit în jurul Moesiei şi au avansat spre NE şi E, toată marginea de nord a Moesiei a suferit progresiv extensie. Această extensie a migrat din vest (Miocen inferior) spre est (Miocen mediu) iar direcţia de întindere apare ca rotită în sens orar, similară cu direcţia de rotaţie a unităţilor Carpatice. Astfel, diferenţele în termeni de evolutie tectonică şi mişcări verticale dintre "foreland"-ul Carpaţilor Orientali şi, respectiv, Meridionali (Platforma Est Europeană/Scitică, respectiv, Moesia) apar ca fiind controlate de proprietăţile reologice (litosferă puternică versus litosferă slabă) precum şi de direcţia de convergenţă/coliziune (coliziune frontală versus coliziune oblică).

CHAPTER 1

INTRODUCTION AND OUTLINE OF THE THESIS

1.1. INTRODUCTION

Foredeep basins develop adjacent to fold-and-thrust belts on the continental foreland plate as it reaches the subduction zone. Subsidence is related to the downward flexure of the foreland in response to orogenic loading (e.g. Beaumont, 1981; Jordan, 1981). Typically, subsidence in foredeep/foreland basins increases towards the orogenic belt and is coeval with its contractional deformations. The amount of subsidence of the flexural foreland basin depends mainly on the orogenic topography and rheology (strength) of the lithosphere. A foreland basin can be bounded on the external side by a fore-bulge, with heights of tens to hundreds of meters (e.g. ~400-500 m Late Cenozoic uplift in northern Oman or Apennines forelands, according to Rodgers and Gunatilaka, 2002 and Billi and Salvini, 2003, respectively). Normal faults roughly parallel to the foreland basin axis can form in the areas where the stress in the foreland plate overcomes locally its strength. Mechanical coupling between the orogenic wedge and the foreland (sensu Ziegler et al., 2002) can (re)activate contractional structures within the latter, sometimes at large distances from the belt and can accentuate the amplitude of potential flexural fore-bulge.

Many orogen/foreland systems “deviate” from the above rules, especially in terms of anomalously high subsidence of the foredeep basin, as well as the position and timing of subsidence with respect to the tectonic evolution of the associated orogenic belt. The Romanian Carpathians provides a natural laboratory for the study of such unusual subsidence and tectonic history of a foreland basin and their relationship with the adjacent fol-and-thrust belt. The maximum depth of the Carpathians Neogene foredeep (up to 13 km) has basically never been explained. Its complexity arises not only from the extreme depth, but also from the timing of subsidence, which partly occurred after the main orogenic contractional event (Mațenco et al., 2003; Bertotti et al., 2003). The site of major subsidence lies in the SE Carpathians Bend region (Focşani Depression) and represents an important topic because of its peculiar location in front of the thrust belt and because it lies next to the area with one of the highest seismicity in Europe (Vrancea Zone).

On the other hand, the foreland of the South-Carpathians and Balkans (i.e. Moesia – Fig. 1.1) keeps the memory of impressive erosion below the Neogene sediments. Widespread, deep canyons (up to a few kilometres-deep and several tens of kilometres-wide) are witness of Paleogene-to-Early Miocene kilometric-scale uplift, and are at odds with a typical foreland evolution. Overall, Moesia has undergone both anomalous uplift and subsidence at various rates and in different areas during the Tertiary, but its response to the displacement of the Carpathians units has been less studied (e.g. Mațenco et al., 1997b; Răbăgia and Mațenco, 1999).

The Carpathians (Fig. 1.1) acquired their arcuate shape during the Tertiary. The Alpine evolution started with Triassic-Early Cretaceous extension, followed by contractional stages until Pleistocene times (e.g. Săndulescu, 1984, 1988). Lying presently on their concave side (in between the South-Carpathians and Balkans – Fig. 1.1), the Moesian foreland played an important role in the shape and tectonic evolution of the Carpathians-Balkans belt (Stille, 1953; Ratschbacher et al., 1993). It is generally accepted that Carpathians units collided with the W-SW part of Moesia in Mid-Cretaceous times, rotated subsequently around its corner and moved towards their present position since the Paleogene (Săndulescu, 1988; Schmid et al.,

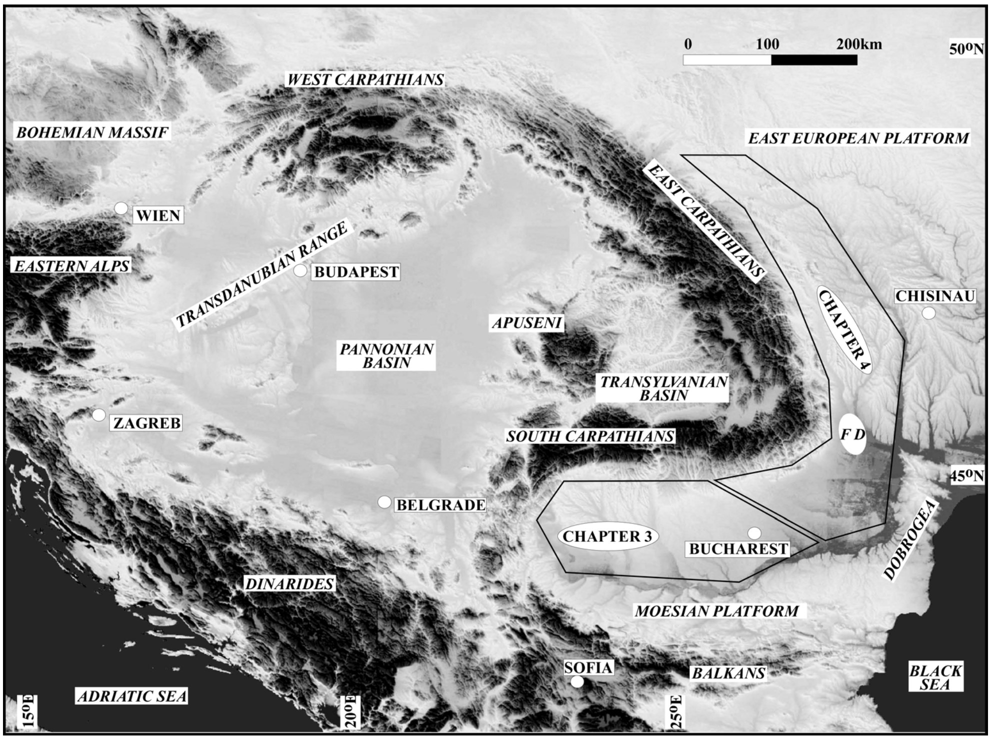

Figure 1.1 *Topography of the Carpathians system. Polygons represent the South- and East-Carpathians foredeep/foreland, which are described in Chapters 3 and 4, respectively. FD represents the approximate position of the Focşani Depression, the deepest zone of the Carpathians foredeep.*

1998; Hippolyte et al., 1999; Mațenco and Schmid, 1999). During the Tertiary, an oceanic or thinned continental lithosphere was subducted below the East-Carpathians (e.g. Săndulescu, 1988; Wortel and Spakman, 2000).

The Carpathians foreland is inhomogeneous, being composed of lithospheric blocks with different crust/mantle thickness, rheology and thermo-tectonic age. Lateral variations in thickness and rheological properties of the foreland plate influence the structural configuration of the orogenic wedge (Mațenco and Bertotti, 2000) and seem to control not only the amount of subsidence recorded by the foreland basins but also the processes responsible for exhumation and volcanism of the Carpathians orogen (e.g. Cloetingh et al., 2004).

A large amount of data has been acquired in the Carpathians foreland since it represents a prolific yet mature petroleum province. Based on well and reflection seismic data (mainly industry seismic lines) the foreland architecture and evolution can be accurately imaged. However, few studies have attempted to integrate the subsidence evolution of the foredeep/foreland with the tectonic history of the orogenic belt. Previous interpretations of the Carpathians foredeep, including the Focşani Depression, (e.g. Gavăt et al., 1966; Gavăt et al., 1969; Cornea and Lăzărescu, 1980; Săndulescu, 1984; Săndulescu and Visarion, 1988; Visarion et al., 1988; Rădulescu, 1988; Dicea, 1995) do not quantitatively describe the basin evolution. Overall, these publications suggested that the Carpathians foreland evolved as a typical flexural foredeep basin during the Tertiary and its structural pattern is solely the result of flexure due to the orogenic loading. However, quantitative studies (Royden and Karner, 1984; Royden, 1993; Mațenco et al., 1997a) indicate that a large discrepancy exists between the amount of subsidence recorded in the foredeep and the magnitude of the orogenic load. As a consequence, “hidden loads” of various origins were invoked in models proposed for the Carpathians region (e.g. Gîrbacea and Frisch, 1998; Wortel and Spakman, 2000; Gvirztman, 2002).

Foreland basin subsidence history and tectonic evolution provide important constraints on the collisional and post-collisional evolution of orogenic belts. Documenting the foredeep/foreland evolution in terms of structure and vertical movements represents also the first step in quantitative modeling of many phenomena associated with mountain building. Numerical modeling of subsidence stages in the Carpathians realm helps us to understand processes acting during and after continental collision in tectonic settings characterized by highly curved plate boundaries.

1.2. SCOPE AND OUTLINE OF THE THESIS

This Thesis focuses on mechanisms controlling the structural style and the pattern of vertical movements in the Carpathians foreland. The syn- and post-collisional tectonic and subsidence history of the foreland is discussed in an attempt to shed light on the processes responsible for the unusual behaviour of the foreland plate. The inferences made can be extrapolated to other fold and thrust belt systems characterized by a complex evolutions and variable lithosphere mechanical properties. Documenting the young tectonic deformations, i.e. those following the main contractional event in the Carpathians orogen (~11 Myr), is of prime importance in integrating lithospheric and surface processes involved in the creation of tectonic topography and natural hazards (Cloetingh et al., 2003). Interpretation of seismic data acquired in the Carpathians foreland allows us to construct detailed images of the progressive changes in the basin shape and structure. The aim is two-fold: first, the spatial and temporal pattern of subsidence is used to develop quantitative models and an alternative explanation to the currently popular models that invoke deep-seated effects of an oceanic/continental mantle slab remnant (see subchapter 2.4). The combination of different tectonic mechanisms can account for the observed pattern of vertical movements, which are otherwise too complicated. Lithospheric memory (Cloetingh and Lankreijer, 2001) has an effect on the lateral change in the foreland plate strength, being the main factor controlling the anomalously large subsidence. The second aim is regionally oriented rather than process-oriented and has potential impact on the petroleum industry. The detailed interpretation of the Romanian Carpathians foreland presented may form a new basis for further exploration and basin analysis studies because the structural evolution shown in this Thesis differs significantly from the “classical opinions”, still advocated.

The organization of the Thesis is outlined in the following. An overview of the Carpathians orogen and its foreland is given in Chapter 2. The evolution of the Carpathians and the characteristics of the various foreland lithospheric blocks are briefly discussed with focus on the Tertiary time span. A special subchapter reviews previously published tectonic models. Since processes controlling the foreland basin evolution are of primary importance in this Thesis, Chapter 2 comprises also a review of the various theoretical models describing the lithospheric reaction to imposed orogenic loads, with a focus on possible mechanisms accounting for ongoing post-thrusting subsidence. Although lithospheric flexure depends on several factors, the lithosphere rheology is the main influence on modeling results.

A detailed interpretation of the Tertiary Carpathians foredeep/foreland is described in Chapters 3 and 4. The descriptions follow the trend of the basin axis, from the South-Carpathians to the East-Carpathians foredeep/foreland (Fig. 1.1). This distinction is made due to significant differences in vertical movements and basin tectonics along the trend. The structural evolution is discussed in successive Tertiary time intervals and is illustrated by relevant interpreted seismic lines. The detailed structural image obtained is used to document the deformation history of the foredeep/foreland and will be further correlated with the tectonics of the Carpathians orogen. In Chapter 3 an analysis is presented of the large-scale erosion that took place in time (~50 Myr) and space on western Moesia and that has no apparent relation with orogenic processes. The transition from regional uplift to subsidence has a direct imprint on the structural and sediment infilling patterns and the study identifies tectonic processes responsible for this apparent large-scale change.

The foredeep/foreland of the East-Carpathians is the topic discussed in Chapter 4 with a focus on the Focşani Depression and the SE Carpathians Bend foreland (Fig. 1.1). This area is spatially juxtaposed with the eastern Moesia/North Dobrogea promontory and partly with the southern part of the East-European/Scythian platform. The basin architecture is discussed for several time intervals. The SE Carpathians Bend Zone is the region where several tectonic scenarios were proposed, but can be rationalised with a quantitative analysis of the basin geometry and subsidence. Lateral variations in subsidence and structural patterns are documented along the East-Carpathians foreland, highlighting differences between the central-northern sector and the southern one. A good correlation has been found to exist between the post-thrusting subsidence evolution in space and the mechanical properties of the lithosphere, i.e. strong East-European/Scythian platform versus weak Moesia.

A quantitative approach of the subsidence in the Carpathians foredeep/foreland is the topic of Chapter 5. The modeling focuses on the SE Carpathians foredeep (Focşani Depression). Subsidence is modeled in two steps, extension and planform flexure, corresponding to the deformation evolution in time. Discrimination between these two subsidence stages and the use of planform instead of simple 2D flexural modeling allow us to develop a novel explanation for the processes responsible for the subsidence of the particular Focşani basin setting. First-order factors producing such unusual subsidence are related mostly to lateral variations in lithospheric strength. The conclusions should be of prime importance for other curved mountain belts overthrusting inhomogeneous forelands.

Chapter 6 integrates the results of the previous chapters at a regional scale and discusses the tectonic history and vertical movements of the Carpathians orogen, hinterland and foreland for several relevant time frames. This regional view indicates that the direction of convergence between the Carpathians upper and foreland plates largely controlled the deformation pattern recorded by the latter. Also, correlation of the vertical movements at a regional scale provides arguments for last-stage lithospheric buckling taking place in the Carpathians realm. It is strongly suggested that basin subsidence is caused by a succession of totally different tectonic processes, rather than a single long-lasting one coupled with deep-seated effects. The last chapter of the Thesis comprises general conclusions and remarks.

CHAPTER 2

OVERVIEW OF THE ROMANIAN CARPATHIANS AND THEIR FORELAND

2.1. INTRODUCTION

The Carpathians forms one of the most arcuate parts of the Alpine chain, changing the orientation from WNW-ENE to NW-SE at the junction between the West- with the East-Carpathians, from NW-SE to E-W in the South-Carpathians and again to E-W in the Balkans via a relatively short N-S segment (Fig. 1.1). This arcuate shape is only partly the result of oroclinal bending (e.g. Schmid et al., 1998). Many recently published kinematic-oriented papers have attempted to provide explanations for this general shape or some particular “inflexion” points, especially the SE Carpathians Bend. These ideas relate mostly to changes in the stress/transport direction of the intra-Carpathians units since the Paleogene (Royden, 1988; Ratschbacher et al., 1993; Linzer, 1996; Morley, 1996; Linzer et al., 1998; Hippolyte et al., 1999; Zweigel et al., 1998; Mațenco and Schmid, 1999; Mațenco and Bertotti, 2000; Gibson, 2001).

The Carpathians foreland (Moesia, North Dobrogea promontory and East-European/Scythian platform) is generally interpreted as comprising a stable lithosphere from Cretaceous onwards, being affected only by flexure in response to the emplacement of the Outer Carpathians nappes in Late Miocene times (e.g. Săndulescu, 1984; Săndulescu and Visarion, 1988; Visarion et al., 1988). A major normal fault system trending parallel to the mountain chain is present (e.g. Paraschiv, 1979a, b; Săndulescu, 1984; Săndulescu and Visarion, 1988; Visarion et al., 1988; Dicea, 1995; Popescu, 1995) and a few transverse crustal-scale faults have been identified as well.

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2.2. CARPATHIANS OROGEN

The Carpathians belt is a result of convergence and ultimately collision between the African and European plates (e.g. Burchfiel, 1976; Săndulescu, 1984, 1988). The Romanian Carpathians comprise a complex Alpine system made up of a succession of thick to thin skinned nappes. The Alpine history is subdivided into an extensional stage, lasting from the Triassic to the Early Cretaceous times, followed by a general contractional stage that began in the Mid-Cretaceous times and ended in the Pleistocene. The latter comprises tectonic units deformed during the Cretaceous (Dacides) emplacing ophiolitic, basement and the most internal cover nappes, as well as tectonic units deformed during the Neogene (Moldavides), as the external thin-skinned nappes (Săndulescu, 1984).

Paleogene, mainly Eocene extensional structures have been documented in the South-Carpathians (e.g. Schmid et al., 1998) and the southern part of the East-Carpathians (Săndulescu, 1992). Based on balanced geological cross-sections through the East-Carpathians, Roure et al. (1993) estimated a shortening amounting to 130 km from the Oligocene onwards.

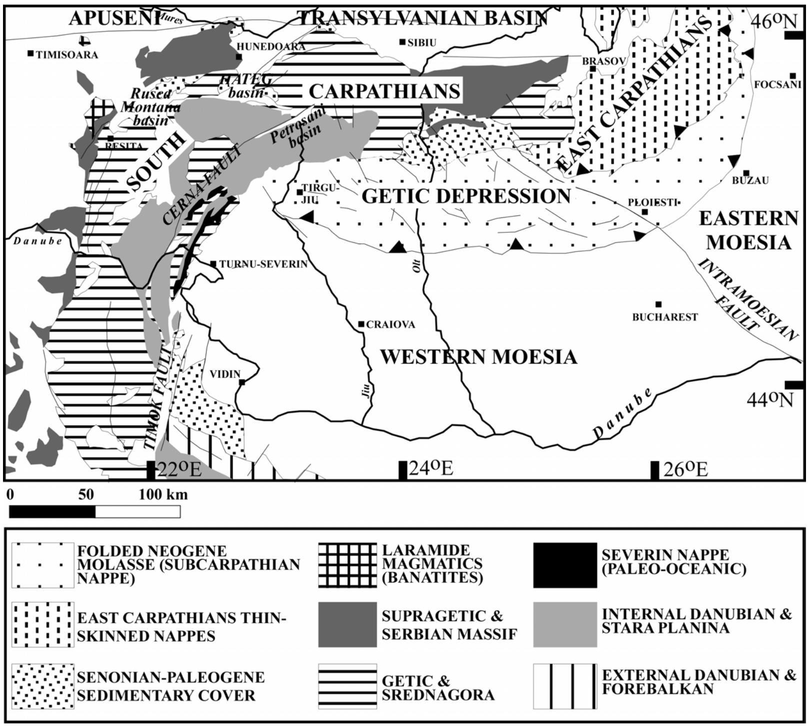


Figure 2.1 *South-Carpathians units. The map is modified from Tari et al. (1997) and references therein. Structures from the South-Carpathians foredeep (namely Getic Depression) are schematically drawn after Răbăgia and Maţenco (1999).*

2.2.1. South-Carpathians

The South-Carpathians (Fig. 2.1) comprise several tectonic units (made up of basement and its sedimentary cover) that underwent Cretaceous contractional deformations during the "Austrian" (Mid-Cretaceous) and "Laramian" (Late Cretaceous) events, sensu Săndulescu (1984, 1988).

A deformed Tertiary foredeep basin developed between the South-Carpathians and the northern part of Moesia (Getic Depression), unconformably covering also the external part of the orogen. Experiencing a rather complex Tertiary evolution (e.g. Dicea and Tomescu, 1969; Răbăgia and Maţenco, 1999), this basin was eventually thrust on top of Moesia during the Late Miocene (Middle Sarmatian). The frontal thrust of this thin-skinned nappe (namely Subcarpathian) is buried below Upper Sarmatian-Quaternary sediments (Dicea and Tomescu, 1969).

2.2.1.1. South-Carpathians nappe pile

Three major thick-skinned nappes have been recognized in the South-Carpathians (Fig. 2.1): Supragetic, Getic and Severin (e.g. Săndulescu, 1984, 1988; Berza et al., 1994). The Supragetic and Getic units are made up of medium-low grade metamorphic basement and a Permian-Middle Cretaceous sedimentary cover. The Severin unit represents remnants of an ocean that had opened during Jurassic-Early Cretaceous times between the Supragetic/Getic units and Moesia (e.g. Săndulescu, 1984). This ocean closed during the Mid-Cretaceous at the same time as the Supragetic nape was thrust on top of the Getic nappe. Contractional deformation resumed in the Late Cretaceous when this nappe stack was emplaced over the Danubian unit ("Autochthonous"), which is usually seen as forming part of Moesia (e.g. Berza et al., 1994). Syn-orogenic basins, such as Rusca Montană and Haţeg (Fig. 2.1), developed in two Late Cretaceous stages on top of the nappe stack: the first stage was syn-contractional (piggy-back basins) while the second one was related to extensional collapse, according to Willingshofer et al. (2001).

Towards the inner part of the South-Carpathians (also stretching into the Balkans and Apuseni Mts.), a magmatic belt ("banatites") formed during the Late Cretaceous (Fig. 2.1; e.g. Săndulescu, 1984). Its origin is still debated, with opinions relating this magmatism either to W-to-SW-wards subduction of the Severin ocean or to N-to-NE-wards consumption of the Vardar ocean (e.g. Ciobanu et al., 2002).

The overall Latest Cretaceous-Tertiary tectonic history is related to docking of the Balkans/Carpathians units on the Moesian margins and a large-scale clockwise rotation of the South-Carpathians around Moesia, from a probable SSW position (Fig. 2.1; e.g. Schmid et al., 1998). After sinistral transpression caused the final thrusting of the Balkans during the Middle Eocene (Doglioni et al., 1996), the South-Carpathians experienced Late Eocene–Early Miocene large-scale orogen-parallel extension (core complex formation), clockwise rotation towards its present position and extensional collapse (e.g. Schmid et al., 1998, Răbăgia and Maţenco, 1999). Significant exhumation took place during the Eocene, as it is documented by FT studies (e.g. Fügenschuh and Schmid, submitted).

The rotational motion of the Carpathians units was achieved mainly during the Oligocene along curved dextral strike-slip faults such as the Timok-Cerna system (Ratschbacher et al., 1993; Fügenschuh and Schmid, submitted). During the Middle Miocene, the South-Carpathians basement nappe pile (Getic/Danubian) was emplaced on top of the northern, distal parts of the Moesian platform (interpretation of profiles in Ştefănescu et al., 1988). The last Late Miocene tectonic episode relates to dextral transpression and thrusting of the South-Carpathians basement units plus a thin-skinned unit developed at the contact (Getic Depression) onto Moesia (Maţenco et al., 1997b). Coevally, uplift of the inner South-Carpathians seems to have occurred, according to Sanders et al. (1999). One should note that in terms of tectonic blocks, Moesia extended as far as north as the Getic/Danubian system until the Late Miocene, when the Getic Depression was peeled off and thrust on top of Moesia.

2.2.1.2. Getic Depression

Lying to the south of South-Carpathians, this Tertiary basin separates the orogenic nappe pile from Moesia (Fig. 2.1). Sediments of the Getic Depression overlie both the orogen and Moesia but the contact between them remains unknown due to the thick sedimentary fill. Originally, the name "Getic Depression" was attributed to a "purely foredeep basin", which evolved from the Paleogene onwards, according to Săndulescu (1984, 1988). However, recent interpretations of a large seismic survey and paleostress/structural analysis within and on the border of the basin (Maţenco et al., 1997b; Răbăgia and Maţenco, 1999) indicate that the tectonic evolution of this area is far from that of a simple foredeep model. Giving these circumstances, the original name Getic Depression becomes loosely defined. Since it has been widely used in the literature, I will retain it to describe only the last foredeep stage, i.e. the syn- and post Late Miocene (Middle Sarmatian) thrusting event.

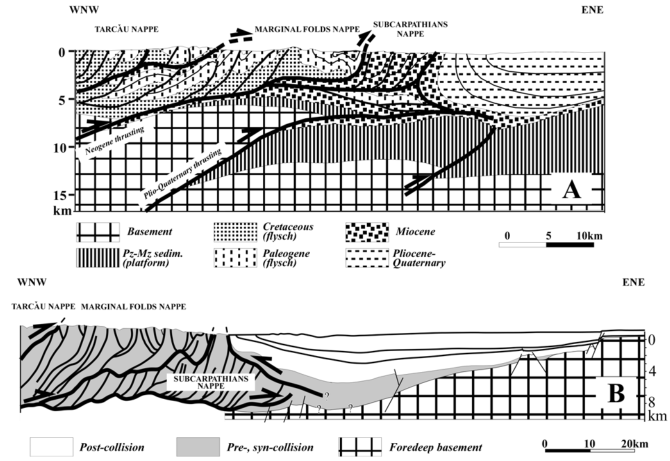


Figure 2.3 (right) *Geological cross-sections through the SE Carpathians Bend zone and Focşani Depression: (**A**) after Roure et al., 1993; (**B**) after Maţenco and Bertotti, 2000.*

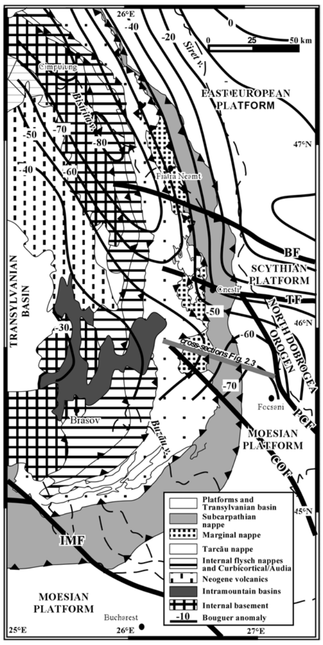


Figure 2.2 (left) *East-Carpathians and the foreland units. The map is modified from Maţenco (1997) and references therein. The numbered lines refer to the Bouguer gravity anomaly (values in mgals). Abbreviations for major faults in the foreland: BF Bistriţa fault; COF Capidava-Ovidiu fault; IMF Intramoesian fault; PCF Peceneaga-Camena fault; TF Trotuş fault.*

The Paleogene kinematics and basin evolution are poorly constrained thanks to overprinting by later deformations. The maximum thickness of the Paleogene deposits is estimated at ~5 km close to the contact with the northern outcropping nappe pile (Jipa, 1980, 1984). Maţenco and Schmid (1999) proposed that the required subsidence was related to a Latest Cretaceous-Paleocene foredeep stage followed by an Eocene-Oligocene extension/transtension event coeval with the core complex formation and Danubian unit unroofing in the South-Carpathians.

The post-Paleogene evolution is separated in: (1) Early Miocene (Burdigalian) extension/transtension along a NW-SE to N-S stress direction; (2) Middle Miocene (late Burdigalian-Badenian) NE-SW oriented contraction; (3) Late Miocene (Middle Sarmatian) dextral transpression, partitioned between NW-SE trending dextral strike-slip faults and N-S oriented thrusting (emplacement of the Subcarpathian nappe) and (4) Latest Miocene (Late Sarmatian-Meotian) strike-slip, with NNE-SSW trending sinistral faults and some reactivated NW-SE trending dextral ones (Maţenco and Schmid, 1999; Răbăgia and Maţenco, 1999). Minor reactivation of the contractional structures occurred at the end of the Pliocene in the eastern part (Maţenco et al., 1997b).

2.2.2. East-Carpathians

In the East-Carpathians, Neogene contractional deformations started with Early Miocene (Early Burdigalian) thrusting of the Curbicortical/Audia nappes, followed by Middle Miocene (Late Burdigalian-Badenian) ~E-W shortening that affected the Tarcău and Marginal

Folds nappes (Maţenco and Bertotti, 2000). Further shortening in a foreland propagating sequence continued during the Late Miocene (Sarmatian) leading to the emplacement of those nappes onto the most external domain, named Subcarpathian. Furthermore, the Subcarpathian nappe was thrust over the foreland (Fig. 2.2). Roure et al. (1993) estimated that 108 km of shortening took place along a cross-section through the SE Carpathians Bend during the Middle-to-Late Miocene (Badenian-to-Sarmatian) time span (Fig. 2.3).

The next deformation stage (Latest Miocene: Latest Sarmatian-Early Meotian) was characterized by a strike-slip regime with NNE-SSW to N-S compressional axes. To the north of Trotuş fault the strike-slip deformation was accommodated by roughly E-W sinistral faults whereas in the southernmost East-Carpathians NW-SE dextral faults were formed. An intermediate transition zone experienced ESE-wards displacement, estimated at 40-50 km according to Maţenco and Bertotti (2000).

In the Bend Zone, Pliocene-Pleistocene out-of-sequence thrusting took place with roughly NNW-SSE-directed shortening (Hippolyte and Săndulescu, 1996) of about 22 km according to Roure et al. (1993) or ~15 km after Maţenco and Bertotti (2000). However, a significant difference exists between these two interpretations: whereas Roure et al. (1993) suggested that this shortening involves blind basement thrusts, Maţenco and Bertotti (2000) considered that only the sedimentary cover was deformed (Figs. 2.3A and B, respectively). Upward movement tilted the most internal parts of the foredeep towards the east, as documented by the presence of marine to brackish sediments at elevations of several hundred meters above sea level. Vertical to steep Late Miocene (Sarmatian-Pontian) strata dip away from the orogenic wedge (Dumitrescu et al., 1970). In fact, the tilting occurs on a large scale (tens of kilometers) and the typical orogenic mechanisms seem to exert apparently little or no control (Cloetingh et al., 2003).

Significant uplift is recorded since Late Badenian-Early Sarmatian times in the central-northern East-Carpathians where >5 km of rocks have been eroded. The main uplift in the Bend Zone is younger and started at the end of the Miocene (Pontian), coeval with the youngest contractions (Sanders et al., 1999).

Towards the inner part of the East-Carpathians, a subduction-related, Neogene calc-alkaline volcanic arc stretches from NW to the SE (Fig. 2.2). The age of magmatic activity becomes progressively younger towards the SE, from ~12 Myr in the NW to ~6 Myr close to the Bend zone (e.g. Mason et al., 1998). At the SE end of this calc-alkaline volcanic arc (in the Perşani Mts.), very young (Quaternary) alkaline magmas erupted from a supposed asthenosphere source (Mason et al., 1998; Chalot-Prat and Gîrbacea, 2000). In between the calc-alkaline and alkaline segments, a transition zone with combined geochemistry comprises volcanic rocks mostly of Pliocene ages.

The SE Carpathians Bend zone represents the site of some of the most intense seismicity in Europe, known as the Vrancea region. The Vrancea earthquakes have an epicentral area of around 30 km x 70 km and their hypocenter distribution describes a nearly vertical column (e.g. Oncescu, 1984; Oncescu et al., 1998; Fig. 2.4A). Most focal mechanisms point to reverse fault geometry with subvertical extension and NW-SE or NE-SW oriented contraction (e.g. Enescu and Enescu, 2000). A significant number of normal and even strike-slip occurs as well (e.g. Enescu and Enescu, 2000). Results of the recent tomography studies (e.g. Sperner et al., 1999; CALIXTO 99 Research Group, 1999; Wortel and Spakman, 2000) reveal a high velocity sub-vertical body containing the earthquakes hypocenters and surrounded by low velocity zones (Fig. 2.4B). Horizontal sections across this body have two different elongations: NE-SW at depths up to 120-130 km and N-S at larger depths (e.g. Sperner et al., 1999).

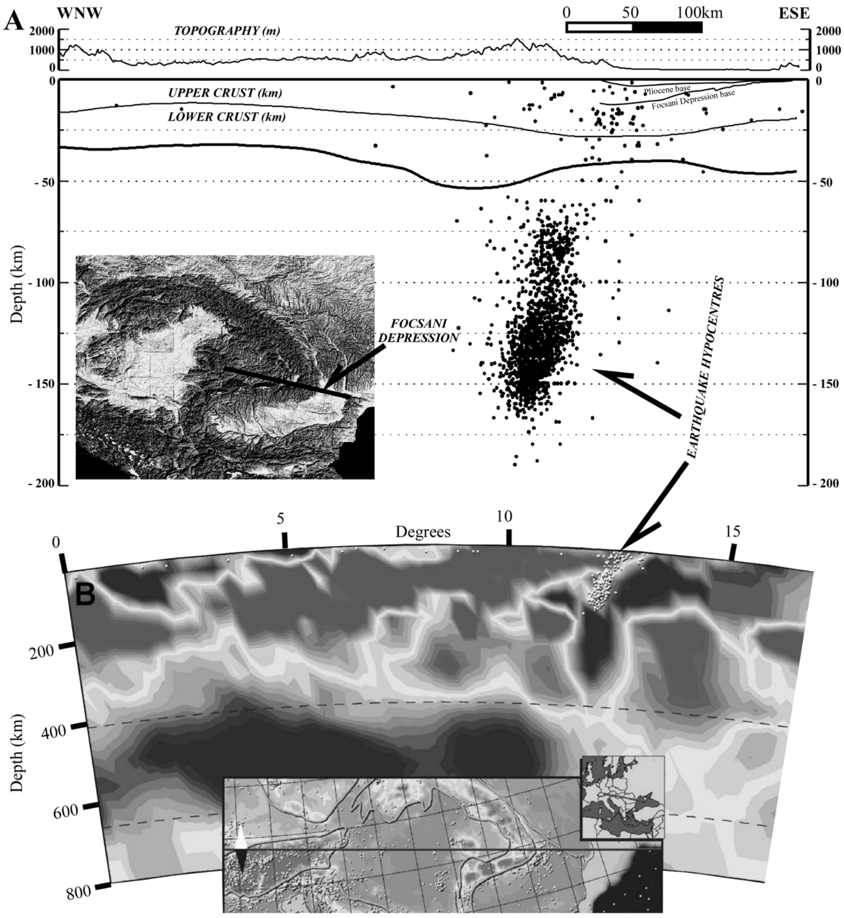


Figure 2.4 *(**A**) Topography, crustal thicknesses (modified from Rădulescu et al., 1976; Enescu et al., 1992), earthquake hypocenter distribution (after Oncescu, 1984; Oncescu et al., 1998) and position of the Focşani Depression (basin and Pliocene base horizons from this Thesis). Location of the cross-section is shown in the inset representing the topography of the Carpathians/Pannonian region. (**B**) Tomographic cross-section through the SE Carpathians (after Wortel and Spakman, 2000). Darker and lighter areas represent higher and lower velocities, respectively. White dots are earthquakes from Vrancea zone. Location of the cross-section is shown in the inset.*

2.3. CARPATHIANS FORELAND

The foreland of the Carpathians is characterized by three tectonic units, platforms in local terminology, made up of crystalline basement overlain by a slightly deformed sedimentary cover (cf. Săndulescu, 1984; Ionesi, 1994): East-European, Scythian and Moesian, respectively (Fig. 2.2). In between Scythian and the Moesian platforms, the North Dobrogea promontory represents the prolongation of the outcropping North Dobrogea orogen west of the Danube. These units are divided by major crustal faults (Fig. 2.2) interpreted from borehole data, gravimetric and magnetometric surveys (e.g. Airinei et al., 1966).

In Romanian literature, the term “platform” is used in different contexts, referring sometimes to the age of the last metamorphism of the crystalline basement and other times to

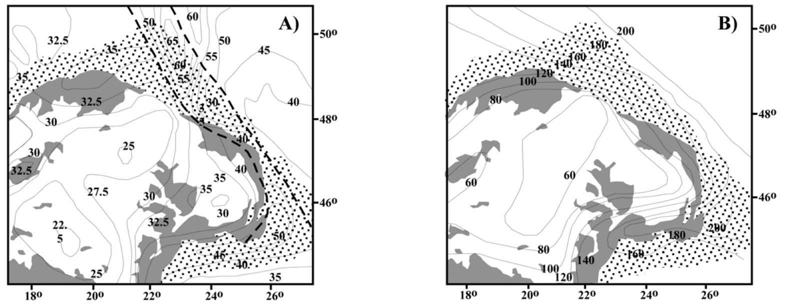


Figure 2.5 *Thickness maps of the crust (**A**) and lithosphere (**B**) of the Carpathians-Pannonian region (after Horváth, 1993). Carpathians inner units (including crystalline basement) are shown schematically as gray areas whereas the outer units and the foredeep are shown as stippled area. In (**A**), the NW-SE trending dash lines represent the margins of TTZ alignment.*

the last major contraction below relatively undeformed sediments. It is not clear if the term refers to the thermo-tectonic age of the stable units. For instance, although Moesia has Precambrian-Early Cambrian metamorphic basement, it is interpreted an "Epi-Hercynian" platform (Săndulescu 1984; Visarion et al., 1988), because its sedimentary cover was affected by the Hercynian deformations and associated magmatism. To add to the confusion, extensional phases are not taken into account in defining the platform age. Accordingly, the term "platform" attributed to Moesia is confusing, as post-Hercynian extension and volcanism are documented (Paraschiv, 1979b; Tari et al., 1997).

There are clear differences in crustal and lithospheric thickness between the various foreland elements (e.g. Rădulescu et al., 1976; Rădulescu, 1988; Săndulescu and Visarion, 1988, Visarion et al., 1988, Horváth, 1993 and references therein; Fig. 2.5), which have significant impact on the Alpine continental subduction and collision in the Carpathians realm (e.g. Cloetingh et al., 2004).

The foreland (East-European/Scythian platform and Moesia) is underthrust by the Subcarpathian nappe. The structural relations between the Carpathians fold-and-thrust belt and foreland are shown in two regional seismic lines (Fig. 2.6; Dicea et al., 1995). In terms of foreland basin depth and timing of the sedimentary fill, the East-European platform and Moesia behaved different, although the emplacement of the Subcarpathian nappe is roughly coeval. These differences will be addressed in detail in the next chapters.

2.3.1. Characteristics of the foreland units

Figure 2.2 shows the tectonic units in the Carpathians foreland. Two WNW-ESE trending faults (Bistrița and Trotuş) separate the Scythian platform from the East-European platform to the north and from the North Dobrogea orogen and Moesia to the south, respectively. The last two units are separated by the NW-SE trending Peceneaga-Camena fault.

In terms of thermo-tectonic ages, it is generally accepted that the East-European platform (the region close to the Carpathians) was formed in the Precambrian and that no subsequent deformations have reset the geotherms. Instead, the last significant deformations within the Scythian platform are related to the Caledonian and Hercynian orogeneses (e.g. Săndulescu, 1984; Săndulescu and Visarion, 1988). These deformations affected also the North Dobrogea orogen (e.g. Săndulescu, 1984; Seghedi, 2001). Moesia is divided in two

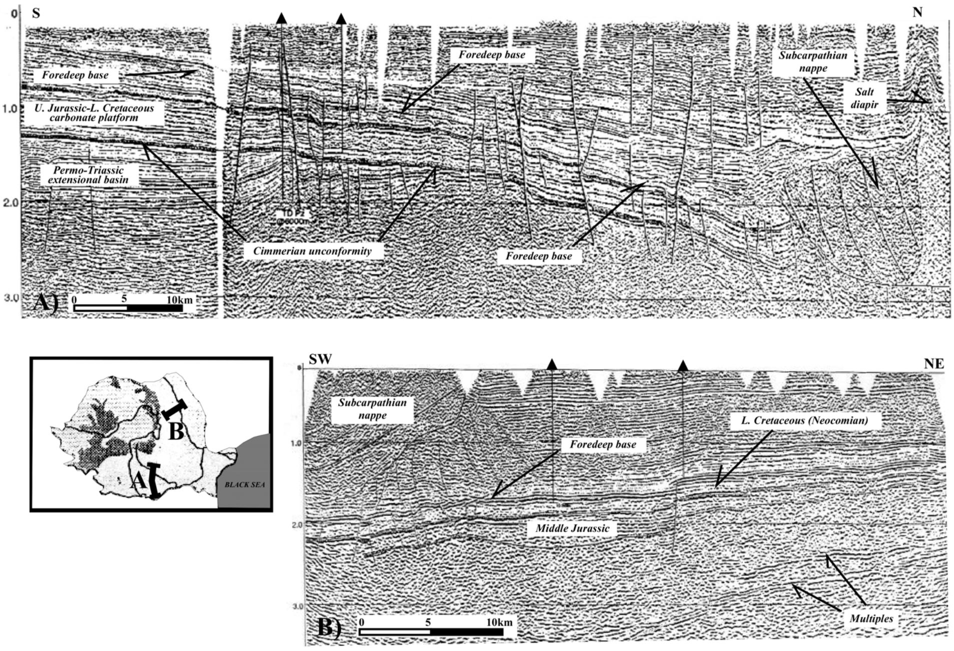


Figure 2.6 *Regional seismic sections showing the structural relations between the Carpathians orogen (Subcarpathian nappe) and foreland (after Dicea et al., 1995).*

blocks by the crustal scale Intramoesian fault (Figs. 2.1 and 2.2) having apparently different thermo-tectonic ages. The eastern part is Precambrian between the Intramoesian and Capidava-Ovidiu faults and Early Cambrian between the Capidava-Ovidiu and Peceneaga-Camena faults (Fig. 2.2). The western part is interpreted as Hercynian as proven by granitic intrusions in the Precambrian metamorphics (Paraschiv, 1979b; Săndulescu, 1984; Visarion et al., 1988).

In terms of crustal and lithosphere thicknesses, the largest values are found in the East European/Scythian and North Dobrogea units where >40 km crustal and >190 km lithosphere thickness is inferred (Rădulescu et al., 1976; Rădulescu, 1988; Enescu et al., 1992; Horváth, 1993 and references therein; Fig. 2.5). Lines of equal thickness parallel the East-Carpathians trend, with the crust thickening towards the orogen and the lithosphere thinning in the same direction. Instead, Moesia has crustal and lithosphere thicknesses ranging between 30-40 km and 150–190 km, respectively (Enescu et al., 1992; Horváth, 1993), both increasing towards the South-Carpathians.

2.3.2. Sedimentary megasequences in the Carpathians foreland

The sedimentary cover of the Carpathians foreland is usually divided into four megasequences separated by major unconformities (e.g. Paraschiv, 1979a, b; Ionesi, 1994; Tari et al., 1997). The first sedimentary megasequence is Paleozoic in age and comprises mainly shallow marine deposits, except for the sequence in the North Dobrogea promontory. The thickness of these sediments ranges between 1–1.5 km on the East-European/Scythian platform to a maximum of 5.5 km on Moesia (e.g. Paraschiv 1979a, b; Ionesi, 1994). In the North Dobrogea promontory, Late Paleozoic flysch and continental sediments followed Early

Paleozoic low-grade metamorphics. Hercynian magmatism is also documented (e.g. Săndulescu, 1984; Ionesi, 1994; Seghedi, 2001). The thickness of the Paleozoic deposits on the North Dobrogea promontory is in order of several hundreds of metres (e.g. Ionesi, 1994).

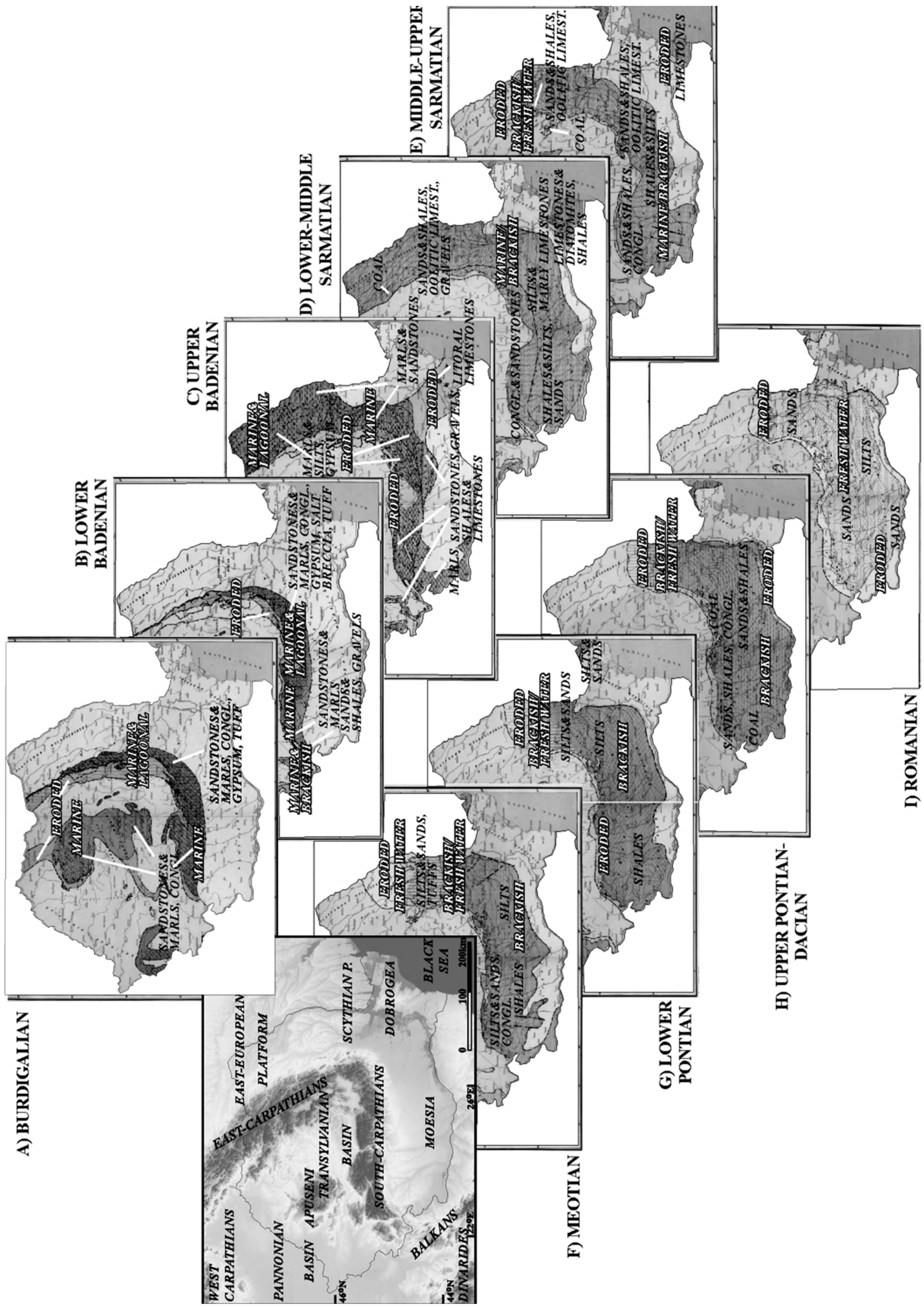
The second sedimentary megasequence follows the unconformity related to the Hercynian deformations and covers the Permian-Triassic time interval. Permo-Triassic sediments range from >1 km-thick on the Scythian platform (e.g. Paraschiv, 1979a) to a maximum of 5 km on Moesia (e.g. Tari et al., 1997). On the East-European platform, this sequence is lacking. Sediments comprise continental to shallow marine clastics, carbonates and local evaporites, as well as intercalations of volcano-clastic layers. It forms a syn- to post-rift sequence deposited in Permian (?)-Middle Triassic extensional basins that were partly inverted during Late Triassic times, well documented in Moesia (Tari et al., 1997; Răbăgia and Tărăpoancă, 1999), Balkans (Georgiev et al., 2001) and the North Dobrogea orogen (Seghedi, 2001). Also, the Scythian platform seems to have evolved as an extensional basin during Permo-Triassic (Seghedi et al., 2003). Sedimentation was accompanied by bimodal volcanism in Moesia (Paraschiv, 1979b), in the Scythian platform (Seghedi et al., 2003) and in the North Dobrogea orogen (e.g. Seghedi, 2001). In the North Dobrogea promontory only thin Triassic deposits can be documented (Ionesi, 1994 and references therein).

The third sedimentary megasequence starts at the end of Liassic times on Moesia and in Dogger times on the Scythian platform with shallow marine clastics deposits. The Malm-Early Cretaceous time span is characterized by platform-type carbonate sedimentation on the entire foreland (e.g. Paraschiv, 1979a, b) except on the North Dobrogea promontory where no Jurassic-Cretaceous deposits can be found (Ionesi, 1994). After an erosional period related to the Mid-Cretaceous "Austrian phase" (Săndulescu, 1984), shallow marine carbonate sedimentation resumed during the Upper Cretaceous time span. The third megasequence terminated almost generally at the end of the Cretaceous as a result of the "Laramian phase" (Săndulescu, 1984). Thin Eocene deposits (tens of meters) are only locally found in SW Moesia and on the East-European platform (Paraschiv, 1979a, b and references therein). The third megasequence has a thickness from 2.4 km on the East-European/Scythian platform to 3.4 km on Moesia (e.g. Ionesi, 1994).

The Carpathians foredeep megasequence contains the fourth package of sediments deposited on the foreland platforms during Neogene times. The basin shows significant lateral variations in thickness and width, which are related to the characteristics of the different basement blocks. This last sedimentary megasequence was deposited after the development of an extensive unconformity that spans almost the entire Paleogene-Lower Miocene time interval. It should be mentioned that in the NW-most and northern parts of Moesia (beneath the Subcarpathian nappe), Paleogene sediments are also found, apparently pinching out to the south (e.g. Dicea and Tomescu, 1969; Dicea, 1995).

At regional scale, the distribution of the Neogene sediments over the Carpathians foreland is shown in Figure 2.7. These lithological maps were constructed by Saulea et al. (1969), who integrated the well database available at that time with information based on field studies.

Figure 2.7 (next page) *Lithofacies maps of the Romanian Carpathians foreland (after Saulea et al., 1969).*


A) BURDIGALIAN
B) LOWER BADENIAN
C) UPPER BADENIAN
D) LOWER-MIDDLE SARMATIAN
E) MIDDLE-UPPER SARMATIAN
F) MEOTIAN
G) LOWER PONTIAN
H) UPPER PONTIAN-DACIAN
I) ROMANIAN
WEST CARPATHIANS
PANNONIAN BASIN
APUSENI
TRANSYLVANIAN BASIN
EAST-CARPATHIANS
SOUTH-CARPATHIANS
EAST-EUROPEAN PLATFORM
SCYTHIAN P.
DOBROGEA
MOESIA
BALKANS
DINARIDES
BLACK SEA

Lower Miocene (Burdigalian) sediments were deposited only in the NW-most part of Moesia close to its contact with the South-Carpathians nappe pile and the Getic Depression where they reach thicknesses of up to 3 km (Răbăgia and Maţenco, 1999 and references therein). During the Paleogene-Early Miocene, the western parts of Moesia were subjected to major erosion (Chapter 3; Paraschiv, 1979b, 1997).

Except for the area with Lower Miocene deposition, this last megasequence commenced during the Middle Miocene (Badenian). In western Moesia, the Badenian sediments infill incised valleys/canyons whereas in its eastern parts they mostly infill extensional basins (see Chapters 3 and 4). Sediments are dominantly clastics in Moesia and clastics interbedded with evaporites on the East-European/Scythian platform and the North Dobrogea promontory. All over the foreland, the Upper Miocene (Sarmatian) deposits consist mainly of clastics with some limestones intervals (e.g. Paraschiv, 1979a, b; Ionesi, 1994). The Uppermost Miocene (Meotian-Pontian) lithofacies are characterized by clastics, which are predominantly siltic-sandy during the Meotian and mainly pelitic-siltic during the Pontian. Starting with the Pliocene, the sedimentary sequence became progressively coarser (e.g. Paraschiv, 1979a, b). Pliocene-Quaternary sedimentation is restricted to the region south of Trotuş fault, i.e. North Dobrogea promontory and Moesia.

The greatest foredeep thickness (~13 km of Badenian-Quaternary deposits) is recorded in the Focşani Depression, which is developed on top of the NE part of Moesia and south of the junction between Moesia, the Scythian platform and the North Dobrogea promontory (Figs. 2.2 and 2.3). The overall thickness of the sedimentary cover in the area of the Focşani Depression is ~18 km, according to e.g. Rădulescu et al. (1976). That means that beneath the Neogene sequence, the older sediments can have thicknesses as great as 5-6 km. The age of these pre-Neogene sediments is, however, unknown, as well as their structural setting. Correlation with the external shallower areas would suggest that both the Paleozoic and Mezozoic megacycles might be present.

In order to summarize the tectono-sedimentary stages and the major deformation events in the Carpathians orogen and foreland, a correlation chart is given in Figure 2.8. The tectonic deformation stages taking place in the orogen are shown starting with the onset of lithospheric stretching during the Triassic, which was followed by break-up and spreading, at least in the Severin ocean.

2.4. PLATE TECTONICS AND FOREDEEP SUBSIDENCE MODELS

2.4.1. Review of previous tectonic models

The evolution of the South- and East-Carpathians is related to the closure of an oceanic (partly thinned continental crust?) basin (Severin ocean and Outer Dacidian Trough, sensu Săndulescu, 1984, 1988) that opened within the European plate during the Jurassic-Early Cretaceous. The Severin domain represents the southern part of this basin, which was eventually obducted during collision of the Apulian and European plates in Late Cretaceous times (e.g. Săndulescu, 1984; Csontos, 1995). The N to NNW-wards displacement of the Apulian promontory continued during Paleogene-Early Miocene and caused the rotation of the Carpathians units around the Moesian corner (Ratschbacher et al., 1993; Csontos, 1995; Schmid et al., 1998). NE-wards movement of the Carpathians units involved the subduction of an oceanic or thinned continental lithosphere below the East-Carpathians units. Ratschbacher et al. (1991) proposed that Apulia indentation in the Eastern Alps during Early Neogene times promoted E-wards escape of crustal units and extension in the Pannonian basin, which was accommodated by equivalent shortening in the East-Carpathians fold-and-thrust belt (also see Royden, 1988, 1993).

Most interpretations assume that the Neogene-Quaternary evolution of the Carpathians is linked to the roll-back of the subducted (oceanic?) slab, which was attached to the East-European plate (e.g. Royden, 1988, 1993; Nemcok et al., 1998; Gîrbacea and Frisch, 1998; Wortel and Spakman, 2000). In most of these models, this slab, at least partly oceanic, becomes eventually detached from the foreland plate (break-off), possibly within its continental part due to collision with the thick and buoyant East-European continental lithosphere (e.g. Wortel and Spakman, 2000).

The models of slab roll-back and subsequently break-off of the oceanic part of lower plate proposed that the direction of slab retreat was either eastwards (Royden, 1988; Gîrbacea and Frisch, 1998) or the detachment of the slab took place progressively along the arc from NW towards SE (Linzer, 1996; Nemcok et al., 1998; Mason et al., 1998; Wortel and Spakman, 2000; Sperner et al., 2001). Some authors (Gîrbacea and Frisch, 1998; Chalot-Prat and Gîrbacea, 2000; Gvirztman, 2002) proposed partial delamination of the continental mantle beneath the external Carpathians after slab detachment and its gradual sinking into the asthenosphere while it is still attached to the foreland lithosphere beneath Vrancea zone. A particular opinion considers Moesia as the plate to which the oceanic slab is attached to instead of East-European platform (Linzer, 1996).

These interpretations infer explanations for the Neogene volcanic arc, the Vrancea seismicity and the amount/timing of subsidence of the Focşani Depression. Accordingly, the Neogene volcanism in the inner part of East-Carpathians gets progressively younger from NW to SE (e.g. Szackacs and Seghedi, 1995; Mason et al., 1998) and also reflects the progression of the slab detachment from the foreland plate. The earthquake hypocenters beneath Vrancea zone (Fig. 2.3) would indicate the (sub)vertical position of the slab, which is also considered the cause of the large Pliocene-Quaternary subsidence in the Focşani Depression (e.g. Chalot-Prat and Gîrbacea, 2000; Wortel and Spakman, 2000).

Recent numerical modeling approaches (Cloetingh et al., 2004) underline the importance of rheological variations in the foreland plate of the East-Carpathians and infer a strong control of differences between the East-European platform and Moesia in the collisional and post-collisional patterns of vertical movements of both the orogen and the foreland. Accordingly, the continental slab lying in the prolongation of the East-European platform keeps its integrity during and post-collision and is subducted in an oceanic-type way due to the high strength of the lithosphere and fast surface processes (erosion) in the orogen. A stable continental subduction is predicted, as well as calk-alkaline magmatism and minor amount of crustal thinning in the backarc region, which corresponds to the Neogene volcanic arc and the Transylvanian basin, respectively. By contrast, to the southern part of the East-Carpathians (Bend zone), a low-strength continental slab is involved in subduction (corresponding to the Moesian lithosphere). Due to the reduced strength and minor role of surface processes at the time of convergence, an unstable subduction is predicted due to development of gravitational Raleigh-Taylor instabilities. Modeling results show that the subduction of the "Moesian" continental slab becomes blocked in the early phases of collision whereas the continental slab sinks in the asthenosphere and is vertically stretched. The predicted foredeep basin reaches depths considerably greater in the southern part of the East-Carpathians than in its central-northern part due to the differences along the arc related to the type of continental subduction and slab behaviour.

Roll-back of a subducted slab was also inferred from 2D flexural modeling studies (Royden and Karner, 1984; Royden, 1988; Maţenco et al., 1997a). According to these papers, the present-day topographic load is inadequate to explain the large thickness of the foredeep, and "hidden loads" are required. Other models link the timing and amount of subsidence in the Focşani Depression to deep-seated phase changes (Artyushkov et al., 1996).

In spite of the above, the presence of an oceanic/continental mantle slab and its impact on the overall tectonic evolution have been questioned, as some of the postulated effects upon structural and subsidence patterns cannot be actually observed in the Carpathians evolution (e.g. Mațenco and Bertotti, 2000; Mațenco et al., 2003; Bertotti et al., 2003).

2.4.2. Post-thrusting subsidence in foredeep basins: theoretical considerations

Numerical modeling has become a widely used approach to develop evolutionary models and the input parameters are of prime importance in determining the accuracy of results. The type of rheology used to simulate the lithospheric response to applied orogenic loads is probably the most sensitive parameter, as it is relatively poorly constrained, particularly for continental plates. Most models assume that a thin elastic plate can approximate the lithosphere and, consequently, that the foreland deflection is found by solving a fourth-order differential equation, which depends on the plate rigidity (or its effective elastic thickness), magnitude of the load applied and intraplate stresses (e.g. Turcotte and Schubert, 1982; Cloetingh, 1988; Ziegler et al., 2002). All published Carpathians flexural models use an elastic approximation (Royden and Karner, 1984; Royden, 1988; Mațenco et al., 1997a). In these models, no post-thrusting subsidence occurs whereas erosion of the orogenic load would lead to isostatic rebound of the foreland plate (rather than its subsidence). Similarly, detachment of the subducted foreland plate slab causes unflexing and thus uplift of the orogenic wedge and its foreland basin.

A different approach assumes viscoelastic rheology for the lithosphere (e.g. Beaumont, 1981), whereby a relaxation of stresses within the lithosphere depends on a time constant, arbitrarily chosen, which ranges from hundreds of thousands to a few millions of years. The viscoelastic behaviour is equivalent to a time-dependent decrease of the plate rigidity. Thus, after progressive orogenic loading terminates, subsidence can continue under the now static load, depending on the time constant controlling the viscoelastic rheology.

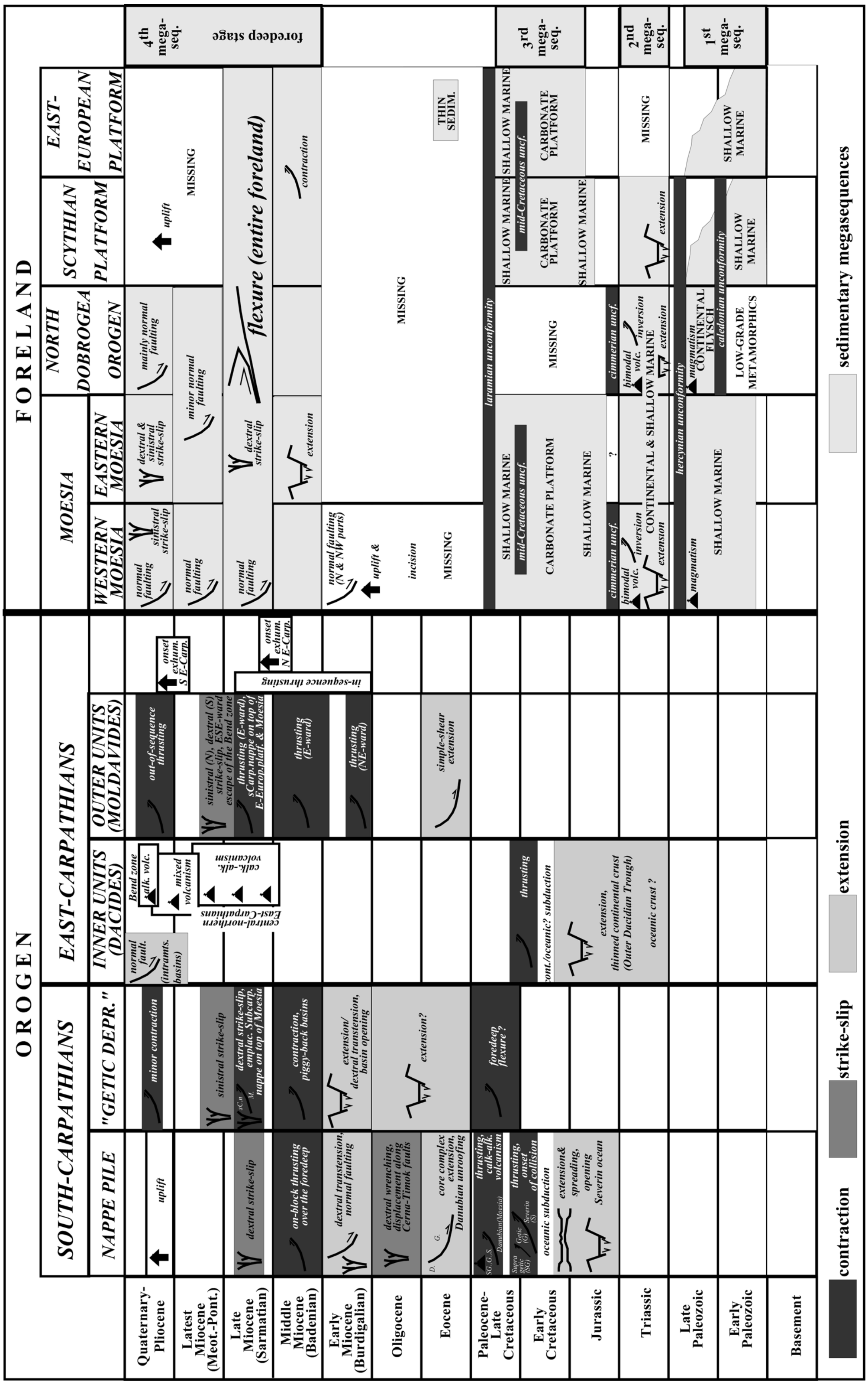
A more realistic approach is given by models using a depth-dependent rheology of the lithosphere, simulating a brittle-elasto-ductile behaviour (e.g. Waschbusch and Royden, 1992; Burov and Diament, 1995). These models start by computing the strength profile of the lithosphere, which is largely controlled by the assumed lithology for the crustal layers and sub-crustal mantle coupled with the thermal field. The effective elastic thickness is given by the integral of the strongest parts of crust and lithospheric mantle (Burov and Diament, 1995). Orogenic loading induces stresses in the lithosphere, which are directly proportional to the curvature of the deflected plate. Extensional bending stresses affect the upper crust of the foreland plate while its lithospheric mantle is in compression. As orogenic loading produced by thrusting (and sometimes by slab pull) increases in time, bending stresses can overcome locally the lithospheric strength profile resulting in inelastic yielding and consequently, the effective elastic thickness of the loaded plate decreases. Thus, at the beginning of loading, deflection of the lithosphere could be almost perfectly elastic but its rigidity progressively decreases due to increasing curvature (bending stresses). Once progressive orogenic loading terminates, the bending stresses related to static orogenic load could be large enough to lower significantly the foreland strength by inelastic yielding and cause post-thrusting subsidence in the foredeep. In addition to the effect of plate curvature upon the strength of lithosphere, Lavier and Steckler (1997) showed that thick sedimentary cover could promote decoupling of the crust from the lithospheric mantle and in turn weakening of the deflected plate.

When the foredeep depth could not be totally accounted for by the magnitude of the orogenic load, usually expressed as the mountain topography, subsurface (or hidden) loads were added, mostly related to slab-pull forces exerted by subducted oceanic or delaminated continental mantle lithosphere (as in the Carpathians case, see previously). However, Stockmal et al. (1986) demonstrated that neglecting paleobathymetry at the time of loading leads to underestimation of the amount of foreland basin subsidence because for a certain period of time thrusting does not create topography.

Mechanical coupling of the orogenic wedge and the foreland lithosphere can cause additional subsidence of the foreland lithosphere and thus "over-deepening" of the foreland basin (Cloetingh, 1988; Cloetingh et al., 1999; Ziegler et al., 2002). On the opposite, mechanical decoupling of the orogenic wedge and the foreland lithosphere causes only load-induced (topographic plus slab load) deflection of the latter (Ziegler et al., 2002).

Another cause responsible for "supplementary" subsidence in foreland basins is related to thermal cooling that follows a rifting event prior the onset of thrusting (e.g. Deségaulx et al., 1991). Depending mainly on the magnitude of extension and the time elapsed until the thrusting proceeds, thermal subsidence in foreland basins may continue after the end of orogenic loading, as seen for instance in Aquitaine basin.

Figure 2.8 (next page) *Tectono-stratigraphic chart summarizing the main deformation events affecting both the Carpathians orogen and its foreland. Also shown, the sedimentary megacycles deposited on the foreland lithospheric blocks.*


OROGEN
FORELAND
SOUTH-CARPATHIANS
EAST-CARPATHIANS
NAPPE PILE
"GETIC DEPR."
INNER UNITS (DACIDES)
OUTER UNITS (MOLDAVIDES)
MOESIA
WESTERN MOESIA
EASTERN MOESIA
NORTH DOBROGEA OROGEN
SCYTHIAN PLATFORM
EAST-EUROPEAN PLATFORM
Quaternary-Pliocene
Latest Miocene (Meot.-Pont.)
Late Miocene (Sarmatian)
Middle Miocene (Badenian)
Early Miocene (Burdigalian)
Oligocene
Eocene
Paleocene-Late Cretaceous
Early Cretaceous
Jurassic
Triassic
Late Paleozoic
Early Paleozoic
Basement
4th mega-seq.
foredeep stage
3rd mega-seq.
2nd mega-seq.
1st mega-seq.
uplift
dextral strike-slip
on-block thrusting over the foredeep
dextral transtension, normal faulting
dextral wrenching, displacement along Cerna-Timok faults
core complex extension, Danubian unroofing
thrusting, calk-alk. volcanism
thrusting, onset of collision
oceanic subduction
extension& spreading, opening Severin ocean
minor contraction
sinistral strike-slip
dextral strike-slip, emplac. Subcarp. nappe on top of Moesia
contraction, piggy-back basins
extension/ dextral transtension, basin opening
extension?
foredeep flexure?
normal fault. (intramts. basins)
Bend zone calk-alk. volc.
mixed volcanism
calk-alk. volcanism
central-northern East-Carpathians
thrusting
cont./oceanic? subduction
extension, thinned continental crust (Outer Dacidian Trough)
oceanic crust ?
out-of-sequence thrusting
sinistral (N), dextral (S) strike-slip, ESE-ward escape of the Bend zone
thrusting (E-ward), sCarp.nappe on top of E-Europ.platf. & Moesia
thrusting (E-ward)
thrusting (NE-ward)
simple-shear extension
in-sequence thrusting
onset exhum. S E-Carp.
onset exhum. N E-Carp.
normal faulting
sinistral strike-slip
dextral & sinistral strike-slip
mainly normal faulting
minor normal faulting
MISSING
flexure (entire foreland)
dextral strike-slip
extension
contraction
normal faulting (N & NW parts)
uplift & incision
THIN SEDIM.
laramian unconformity
SHALLOW MARINE
mid-Cretaceous unf.
CARBONATE PLATFORM
cimmerian unf.
bimodal volc.
inversion
CONTINENTAL & SHALLOW MARINE
hercynian unconformity
magmatism
CONTINENTAL FLYSCH
caledonian unconformity
LOW-GRADE METAMORPHICS
contraction
strike-slip
extension
sedimentary megasequences

CHAPTER 3

FROM REGIONAL UPLIFT TO REGIONAL SUBSIDENCE IN FORELAND BASINS: THE CASE OF THE TERTIARY SOUTH-CARPATHIANS - BALKANS FORELAND (WESTERN MOESIA)

3.1. INTRODUCTION

In western Moesia, i.e. west of the Intramoesian fault, the Late Alpine (Latest Cretaceous-Tertiary) tectonic history is characterized by the docking of the Balkans/Carpathians at the Moesian margins and transport of the South-Carpathians around its western corner (Fig. 3.1A). This clockwise rotation was accompanied in the South-Carpathians by large-scale orogen parallel extension, core-complex formation and extensional collapse during the Eocene–Early Miocene (Schmid et al., 1998; Maţenco and Schmid, 1999; Fügenschuh and Schmid, submitted). Part of the displacement was accommodated in Oligocene times by dextral translation along the curved Cerna-Timok fault system (Fig. 3.1A; Ratschbacher et al., 1993; Berza et al., 1994).

Two Tertiary foredeeps are evident in western Moesia, which are associated with opposite-polarity orogenic wedges (Fig. 3.1B). The southern foredeep formed during the Eocene evolution of the Balkans thrust belt (Doglioni et al., 1996; Georgiev et al., 2001) while the northern formed during the Late Miocene thrusting of the South-Carpathians/Getic Depression belt (Săndulescu, 1984, 1988; Ştefănescu et al., 1988; Maţenco et al., 1997b).

The most interesting feature in western Moesia is an extensive system of incised valleys/canyons that involved kilometers-scale erosion, which is contemporaneous and post-dates the Latest Cretaceous-Eocene emplacement of the Carpathians/Balkans nappes and pre-dates the Miocene docking of the South-Carpathians units. Generally, forelands record only a limited amount of erosion during thrust emplacement, associated with fore-bulge flexural uplift. The magnitude of this uplift is normally in the order of tens up to a few hundreds of metres.

In contrast, kilometers-deep erosion in foreland basins during ongoing orogenic deformation is seldom observed, because these canyons imply regional major uplift instead of regional subsidence. Uplifts in order of 1-3 km (e.g. Bohemian massif or Polish anticlinorium) indicate intraplate compressional deformation rather than flexural fore-bulge alone, which is the effect of strong mechanical coupling between the orogenic wedges and the forelands (Ziegler et al., 2002).

The erosion is roughly coeval with extension/transtension along the northern margin of Moesia (e.g. Răbăgia and Maţenco, 1999) and in the South-Carpathians nappe stack (e.g. Maţenco and Schmid, 1999; Fügenschuh and Schmid, submitted). The end of extension and the onset of contractional deformation in the Getic Depression coincide with the switch from erosion to deposition stage in western Moesia. Regional subsidence in the latter appears to be associated with the Late Miocene emplacement of the Subcarpathian nappe and continued after this thrusting event. The subsidence post-dating thrusting is associated with important positive vertical movements in the orogen, at least during the last stages, as is proven by the presence of Middle Miocene (Badenian) shallow marine sedimentary remnants at currently elevations of ~1.5 km (e.g. Maţenco et al., 1997b).

This chapter is partly based on: Tărăpoancă, M., L. Maţenco, G. Bertotti, J. Matreşu, S.A.P.L. Cloetingh and C. Dinu, Deep erosion in foreland basins: the case of Moesian Platform; Geology, submitted.

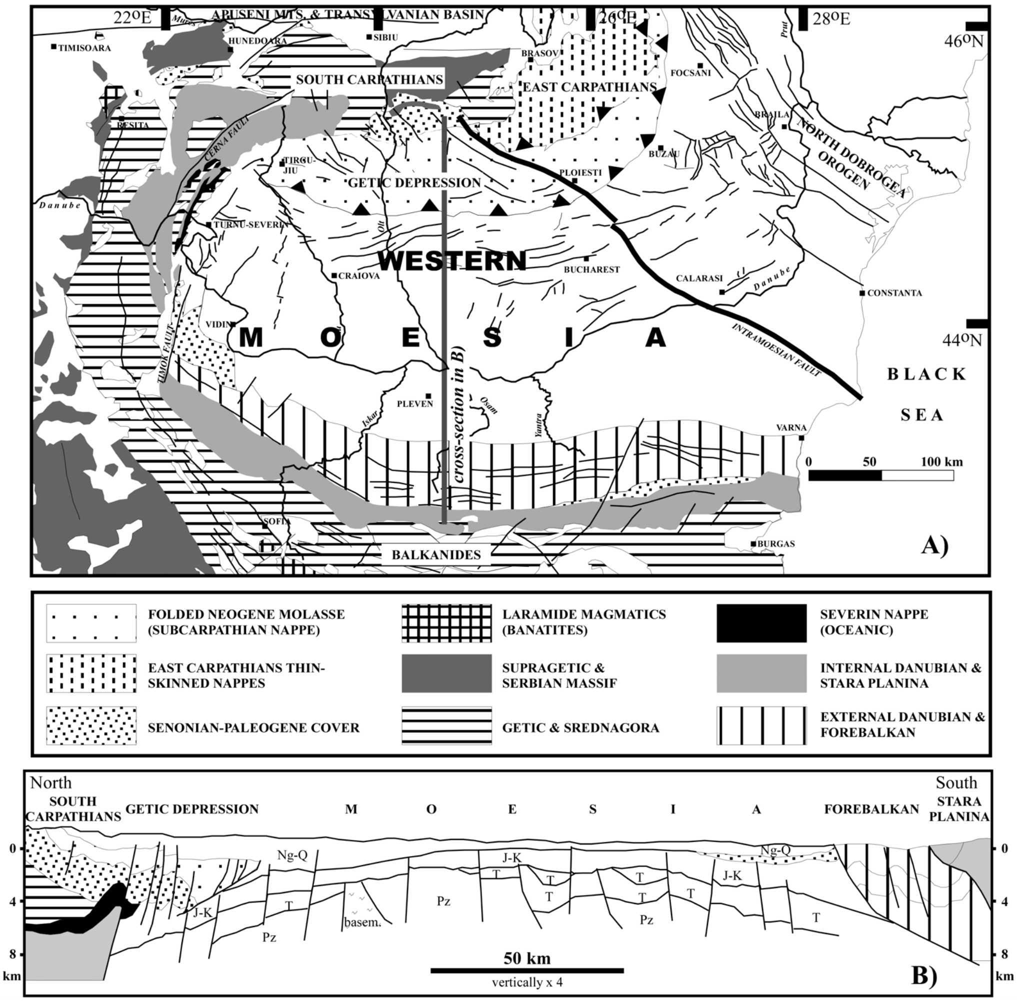
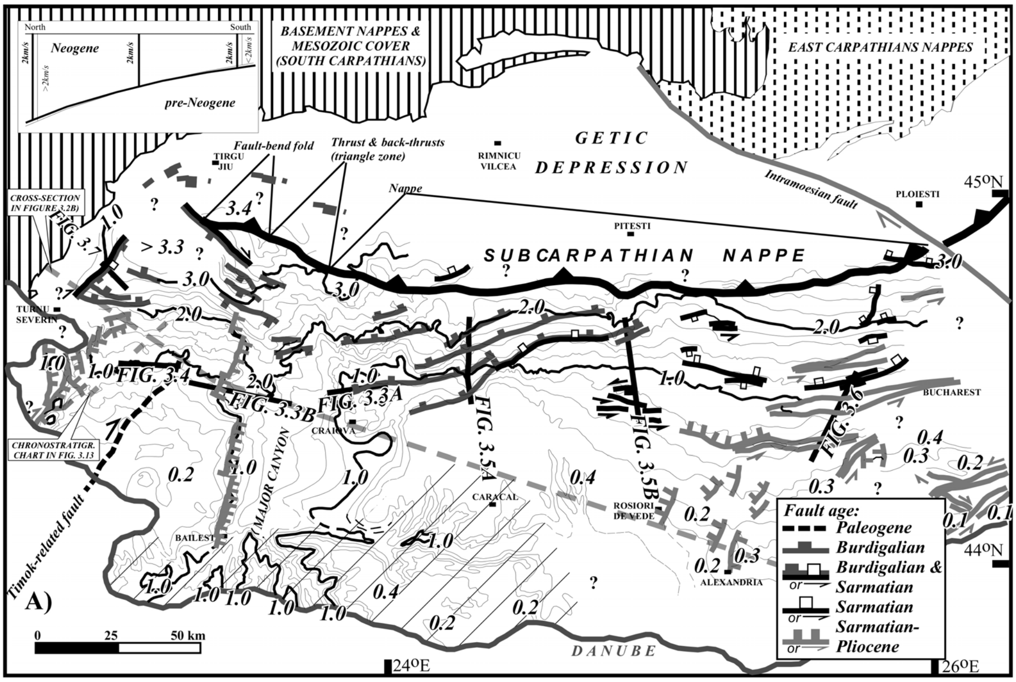
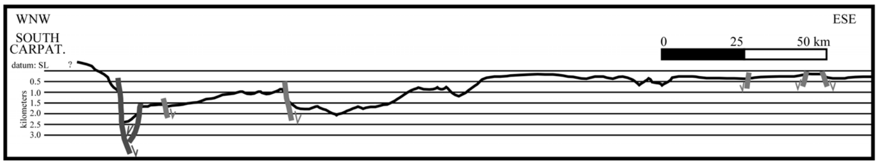


Figure 3.1 (**A**) *Location of Moesia between the Carpathians to the north and west and the Balkanides to the south (modified after Tari et al., 1997 and references therein). The Intramoesian fault represents the boundary between western and eastern Moesia (structural pattern from this Thesis). Structures in the Getic Depression are simplified after Răbăgia and Maţenco (1999).* (**B**) *Schematic N-S cross-section across the South-Carpathians, Moesia and Balkans (modified after Bergerat and Pironkov, 1994; Tari et al., 1997). Note the various foredeep stages of Moesia: Cretaceous and Paleogene to the south and Neogene to the north. Ng-Q: Neogene-Quaternary; J-K: Jurassic-Cretaceous; T: Triassic; Pz: Paleozoic.*

As the processes responsible for these successive and contrasting vertical movements are of prime importance for the foreland evolution, it is necessary to describe the morphological and structural features of both the erosional and subsidence stage. The proposed geometry relies on the interpretation that we have done using a large seismic survey (a few thousands of kilometres) in the western Moesia. From this survey, a map of the major erosional unconformity representing the base of Neogene has been derived (Fig. 3.2A).

For the southernmost area where the seismic coverage is old and scarce (hatched region in Figure 3.2A), we have integrated the interpretation of Paraschiv (1997), which is mostly based on well data.





B)

Figure 3.2 (**A**) *Structural map of the pre-Neogene unconformity (in kilometers below sea level). In the areas north of the frontal Subcarpathian thrust, contours refer to the base of the autochthonous Neogene deposits. In the NW, the dashed lines represent normal faults from Răbăgia and Maţenco (1999). Continuous sedimentation replaces erosion in the N and NW. The conversion from TWT time to depth was made using a constant velocity of 2 km/s for the Neogene sequence, based on the average of a large well check-shots database. The map underestimates the depth in the northern areas of high sediment thickness and overestimates it in the low thickness areas in the south (inset in the upper-left corner). In the southernmost part of the map (hatched region), the contours are taken from Paraschiv (1997). Thick black lines represent seismic lines with their corresponding figure number.*
(**B**) *WNW-ESE cross-section through western-central Moesia showing the pre-Neogene unconformity and its erosional relief. Location in A).*

3.2. PRESENT-DAY SETTING

The most apparent large-scale feature in western Moesia (Fig. 3.2A) is a major erosional unconformity, which developed during Paleogene–Early Miocene (~50 Myr) times and is referred to by Paraschiv (1997) as the “pre-Paratethys denudational surface”. The morphology of this surface can be spectacular, with major valleys/canyons deeply incising (up to 2 km, Fig. 3.2B) the Mesozoic and locally the Paleozoic sequence (Paraschiv, 1979b). These erosional features are presently buried beneath younger sediments, being additionally distorted by slight Middle? -Late Miocene to Quaternary deformations. As shown in Figure 3.2A, the erosional surface morphology indicates a NW-SE to N-S directed valley network that is evident especially in the western part of Moesia. Overall, this morphological surface dips roughly to the north and most of the valleys apparently deepen in the same direction. The most impressive erosional feature is the >100 km-long canyon in the western part of Moesia. Its roughly straight segment reaches 40 km-width and >1.5 km-depth. Close to the Danube, the canyon splits into three narrower valleys, which deepen towards the Balkans. The prolongation of the canyon to the south of the Danube can be correlated with the incised valleys interpreted by Vuchev et al. (1994).

Between the regions of northwards- and southwards-dipping canyons, a large-scale WNW-ESE oriented shallow area stretches across the central-southern part of Moesia, from Craiova to Alexandria (Fig. 3.2A). To the south, some N-S incised valleys several kilometres wide and a few hundreds metres deep connect to an E-W oriented tributary of the major canyon. This tributary is up to 1km-deep and ~10 km-wide. The easternmost N-S oriented valleys apparently deepen towards the Balkans.

Towards the north and especially NW, the erosional surface disappears and a section of fairly continuous Eocene-Lower Miocene sediments is developed. This is the area where the base of the autochthonous Neogene sequence reaches the greatest depths (minimum 3.5 km) and where extensional/transtensional basins are interpreted to have opened during the Eocene (?)-Early Miocene (Răbăgia and Maţenco, 1999). Presently, most of the Eocene-Early Miocene deposits are deformed and incorporated in the Subcarpathian nappe, emplaced in the Late Miocene, except in the NW-most part of Moesia where the thrusting was transferred to strike-slip deformations (Fig. 3.2A).

A NNE-SSW trending fault system lies close to the contact of the westernmost Moesia with the outcropping South Carpathians nappe pile (Fig. 3.2A). In the hanging-wall of the westernmost faults, the amount of downwards displacement of Moesia increases from a few hundreds of meters in the south to >1.5 km in the north. The contact between western Moesia and the South Carpathians has been tentatively considered to be the NE-wards prolongation of the Timok strike-slip fault (e.g. Visarion et al., 1988; Fig. 3.1A). This assumption is based on the fact that the upper South Carpathians thick-skinned nappe (Getic nappe, see Fig. 3.1A) cannot be observed below the Moesian Neogene sequence, which simply onlaps the nappe pile.

Another NNE-SSW trending fault system is documented along the western margin of the major canyon (shown in light gray in Figures 3.2A and B). One could speculate that a genetic relationship exists between this normal fault system with a couple of hundreds of metres offset and the canyon morphology. However, careful observation shows that these faults have a Late Miocene (Sarmatian)-Pliocene age, clearly post-dating the erosional period (see next section). The region to the east of the major canyon is affected mainly by ENE-WSW trending fault systems (Fig. 3.2A). Here, faulting occurred during different Tertiary time periods. Consequently, faults older than the onset of sedimentation (Badenian) and that were not subsequently reactivated do not offset the map contours.

3.3. EROSIONAL STAGE
3.3.1. Lateral variation of the amount of erosion

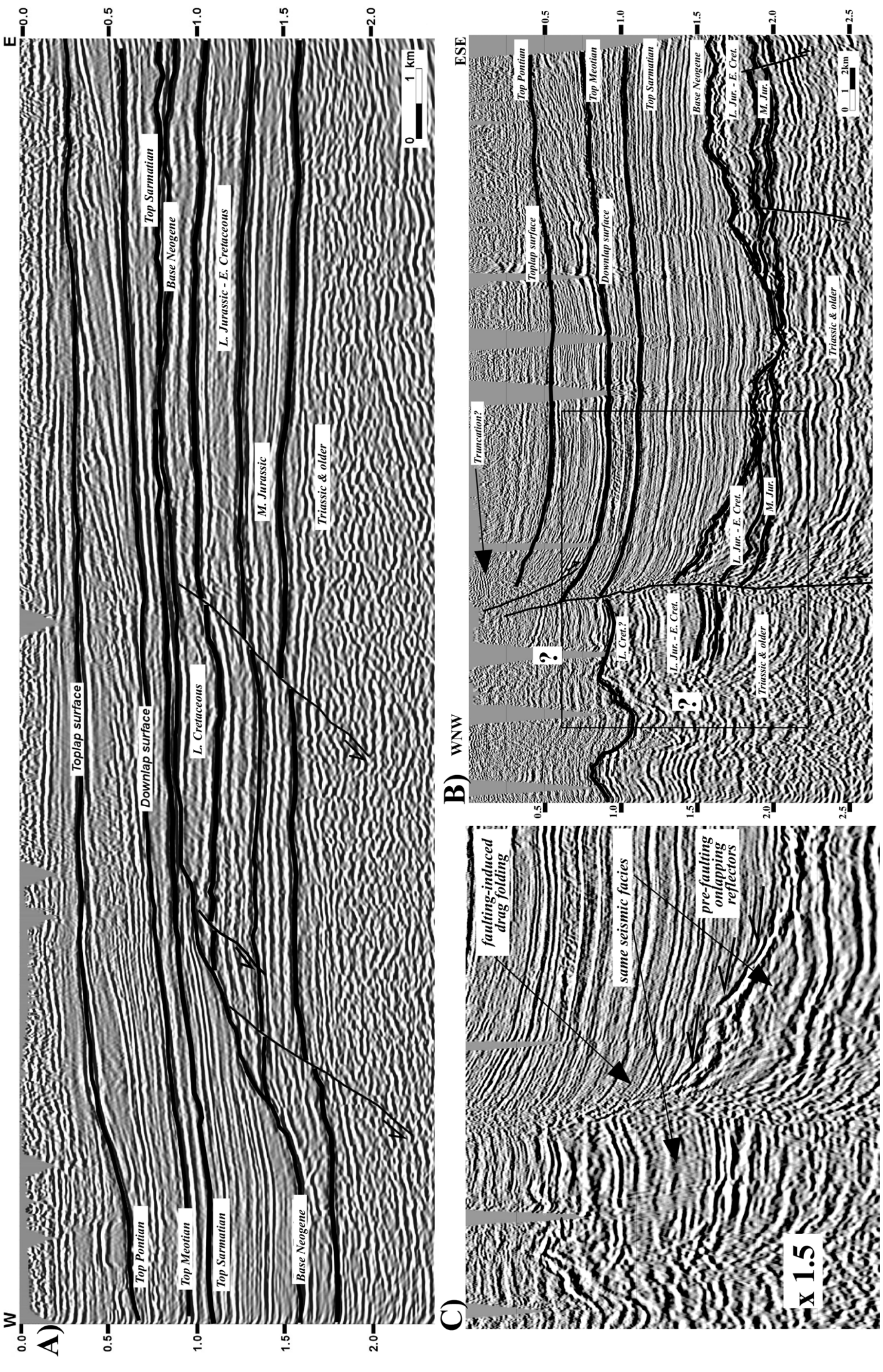
The depth of truncation increases generally from the east to the west where the largest canyon is observed (Figs. 3.2A and B). In the northern areas, the major canyon cut down into strata as old as the Triassic and locally Paleozoic (Paraschiv, 1979b, 1997). The eastern margin of the main canyon is shown in Figure 3.3A, where erosion cut down ~0.6 s (600-700m), removing the entire Upper Jurassic-Cretaceous sequence and part of the Middle Jurassic one. At least two flat segments are present on the erosional unconformity and may be speculated to represent paleo-terraces. The canyon fill consists mostly of Upper Miocene (Sarmatian) sediments, with basal Middle Miocene (Badenian) deposits (e.g. Chişcan et al., 1980).

The western margin of this very spectacular canyon is presented in Figure 3.3B. In contrast to the eastern margin, the western one is normal faulted, clearly at a moment post-dating the erosional unconformity (Figs. 3.3B and C). The sedimentary fill of the canyon do not show any evidence for syn-faulting growth or tilting. Instead, the basal reflectors have bi-directional onlap terminations, typical for incised valleys fill. In the footwall of the normal fault, another 2 km-wide erosional channel may be interpreted (Fig. 3.3B).

Westward of the major canyon, the erosional surface is documented in the Figure 3.4, where two main paleo-valleys with ~1 km width/~0.1-0.2 s (~150 m) depth and few smaller erosional features are interpreted to be present. Along the valleys, the erosion removed Upper Cretaceous deposits, the sedimentary fill being Middle? -Upper Miocene in age (e.g. Paraschiv, 1979b, 1997). Figure 3.4 indicates that the observed depth-difference between the bottom of the major canyon and its margins (e.g. Fig. 3.2B) provides only the lowest estimates of the total amount of erosion.

In the central-eastern part, along two N-S oriented profiles (Fig. 3.5A and B) the same unconformity is observed, with the amount of truncation increasing N-wards, as indicated by the progressive removal of Cretaceous sequences. This pattern is at odds with

Figure 3.3 (next page) (**A**) *Interpreted seismic line across the eastern margin of the major canyon. Location in Figure 3.2A. Note the deep incision in the western part of the line. Minor normal faults clearly die out at the erosional unconformity. The oldest sediments onlapping the canyon margin are Badenian in age. By the end of the Sarmatian, the erosional relief became completely buried. A spectacular W-wards prograding seismic configuration characterizes the Pontian sedimentats. Oblique-tangential seismic facies indicates a relatively high-energy setting. Vertical scale is two-way travel time.*
(**B**) *Interpreted seismic line across the western margin of the major canyon. Location in Figure 3.2A. Except for the spectacular erosional relief at the base Neogene, note the normal fault offseting the unconformity with ~0.5 s (~500-600 m). Based on the parallel seismic facies of the sediments that fill the canyon and the termination of reflectors that simply onlap the hanging-wall (fault zone is magnified in* **C***), we infer that the fault is younger than the erosional unconformity. Onset of faulting could be in the Sarmatian, but we cannot be more precise due to the poor quality of the seismic information in the footwall at shallower levels. The same prograding seismic facies characterizes the Pontian sequence. The clinoform configuration indicates a water depth of ~300 m (~300ms). This deepening of the basin floor can be related to movements along the normal fault (in* **C***, also note drag folding in the hanging-wall) than to sediment compaction.*


A)
W
E
Toplap surface
Downlap surface
Top Pontian
Top Meotian
Top Sarmatian
Base Neogene
L. Cretaceous
L. Jurassic - E. Cretaceous
M. Jurassic
Triassic & older
0 1 km
B)
WNW
ESE
Top Pontian
Top Meotian
Top Sarmatian
Base Neogene
Toplap surface
Downlap surface
Truncation?
L. Cret.?
L. Jur. - E. Cret.
M. Jur.
Triassic & older
0 1 2km
C)
faulting-induced drag folding
same seismic facies
pre-faulting onlapping reflectors
x 1.5

theoretical foredeep models where the main erosion is expected in the flexural-bulge area, i.e. to the south in our case. This suggests that the northern part of Moesia had a higher elevation than the southern one before the Late Miocene flexure. As in the western area, the first onlaps are of Middle Miocene (Badenian) age, the main fill being Upper Miocene (Sarmatian) in age. Farther to the east, the erosion at the base Neogene is minor (Fig. 3.6). Here, the incised valleys are smaller than the seismic resolution (~20-30 m). The continental, partly coarse lag sediments that were often intercepted by wells above the Cretaceous sequence (e.g. Paraschiv, 1979b and references therein) suggest the presence of such valleys.

The large-scale erosional features and the presence of continental deposits above the regional unconformity indicate that a river network was involved in creating this morphology. The magnitude of erosion (kilometres-scale) implies Moesia experienced major uplift in order to allow the rivers to incise deeply. At this stage, we are not able to be more precise about the amount of this uplift or to determine whether a regional sea level drop may have contributed to the observed incision.

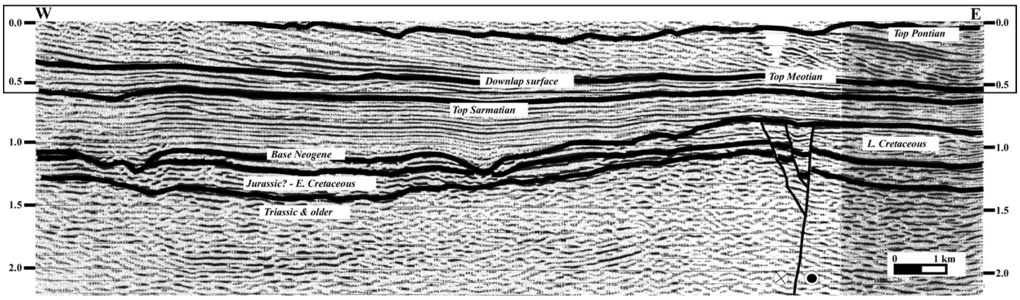
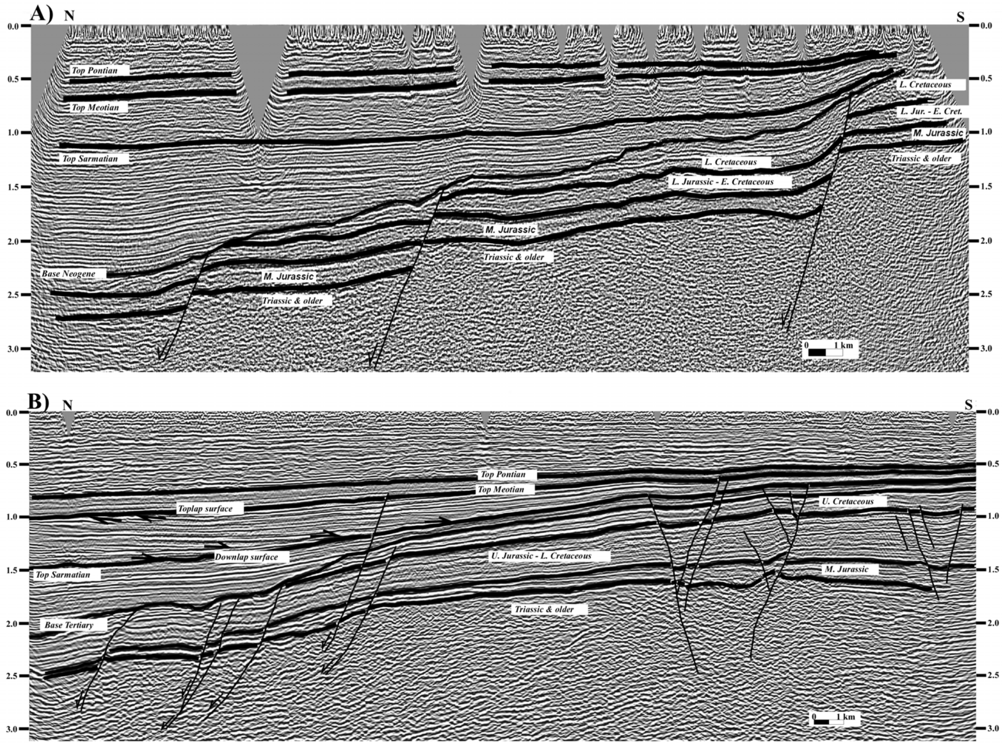


Figure 3.4 *Interpreted seismic line from the western part of Moesia. Location in Figure 3.2A. Note the erosional relief and the incised valleys at the base Neogene. Parallel-seismic facies characterizes the Sarmatian and Meotian sequences. The Sarmatian sequence thickens W-wards (onlaps E-wards). A dramatic change occurred at the Meotian/Pontian boundary above which spectacular E-wards deltaic prograding bodies can be seen. The top of the Pontian sequence is represented by an erosional unconformity. The shallow part of the line marked by a long rectangle is interpreted in greater detail in Figure 3.12. Vertical scale is two-way travel time.*

Figure 3.5 (next page) *Interpreted N-S oriented seismic lines in the central-eastern part. Location in Figure 3.2A. Note the well-developed erosional unconformity separating the Neogene sequence from the Mesozoic one. The unconformity dips to the north and the Late Miocene (Badenian to Sarmatian) sequence has a typical foredeep wedge-shape in both lines. The Pliocene sequence also thickens towards north but more gently than the Sarmatian one. However, in contrast to typical foredeep models that show maximum erosion towards the fore-bulge (to the south in this case), the opposite is observed in these lines: in* ***A****) and* ***B****) the amount of erosion increases N-wards as evidenced by the progressively truncation of Mesozoic sequences. Normal faults younger than Cretaceous dying out at the erosional unconformity can be observed in* ***A****) and in the northern part of* ***B****). In* ***A****) the maximum vertical offset is ~1 km, however, no syn-rift sequence can be identified. At the southern end of line* ***B****) a flower structure is interpreted (note both normal and reverse separations). Vertical scale is two-way travel time.*

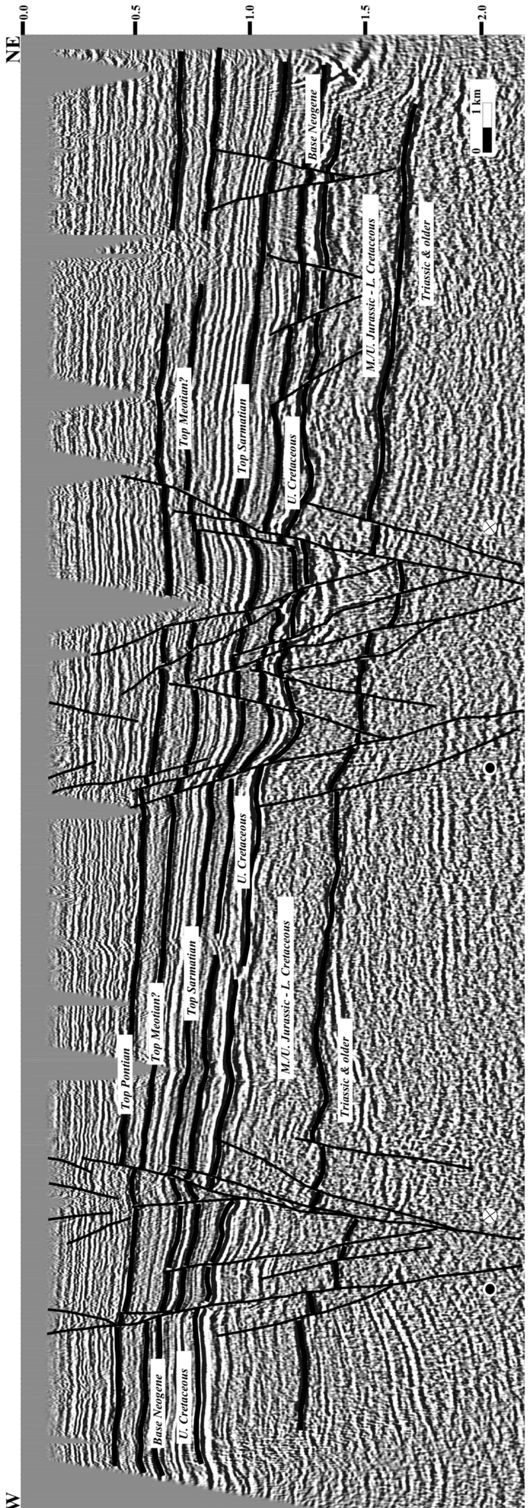


3.3.2. Syn-erosion tectonic deformations

Surprisingly, the Paleogene-Early Miocene erosion cannot be entirely related to any of the contractional phases in the overall tectonic framework (Fig. 3.2A). Inferences about deformations that occurred during the erosional period are derived from faults that affect the Upper Cretaceous sequence and that are truncated by the base Neogene unconformity, such as the negative flower structure seen on in Figure 3.4. Because this fault trends parallel to the Cerna-Timok strike-slip fault system (Figs. 3.1A and 3.2A), a dextral sense of displacement and an Oligocene age are proposed (a description of the Cerna-Timok faults tectonic activity can be found in Berza and Drăgănescu, 1988; Ratschbacher et al., 1993; Berza et al., 1994). The westernmost younger fault system (yellow in Figure 3.2A) might represent the prolongation of Timok fault as previously proposed by Visarion et al. (1988), but no indication of a strike-slip movement prior to the Neogene can be documented.

Normal faults with vertical offsets up to 1 km are best observed in Figure 3.5A (which correlate with faults interpreted in Figures 3.3 and 3.5B). No syn-rift sequence is

Figure 3.6 (next page) *Interpreted seismic line in the eastern part. Location in Figure 3.2A. Note that only minor erosion occurs at the base Neogene unconformity. Thin Middle Miocene (Badenian) continental deposits were documented (e.g. Paraschiv, 1979b) overlying the unconformity. Significant young faulting took place. Differences in layer thicknesses across the two major fault systems as well as the typical negative flower structures attest to a strike-slip type of deformation. It can be observed that all Neogene sequences thicken to the north. Vertical scale is two-way travel time.*


NE
SW
0.0
0.5
1.0
1.5
2.0
Base Neogene
M./U. Jurassic - L. Cretaceous
Triassic & older
Top Meotian?
Top Sarmatian
U. Cretaceous
Top Pontian
0
1 km

present in the interpreted profiles in spite of the significant throws of these faults (e.g. Fig. 3.5A). Correspondingly faulting affected an uplifted and subaerially exposed area. Most of these normal faults strike ENE-WSW with some local NW-SE trending ones (dark gray faults in Figure 3.2A). They appear to sub parallel the extensional/transtensional structures of the present Getic Depression described by Răbăgia and Maţenco (1999). One of the largest faults at the westernmost margin of the basin is shown in Figure 3.7. The pattern of the Lower Miocene (Burdigalian) sequence documents syn-tectonic deposition. As a result, we propose the same Early Miocene (Burdigalian) age for the normal faulting occurring southwards (e.g. Fig. 3.5A).

Eocene orogen-parallel extension in the South-Carpathians (e.g. Fügenschuh and Schmid, submitted) that was roughly coeval with the N-wards thrusting of the Balkans (Doglioni et al., 1996) was transferred at a later stage to the northern part of Moesia, where normal faults affect a larger area than previously documented (Răbăgia and Maţenco, 1999). The Burdigalian extension affecting the South-Carpathians nappes produced faults with vertical offset as large as 3.5–5km, as derived from FT studies (Fügenschuh and Schmid, submitted).

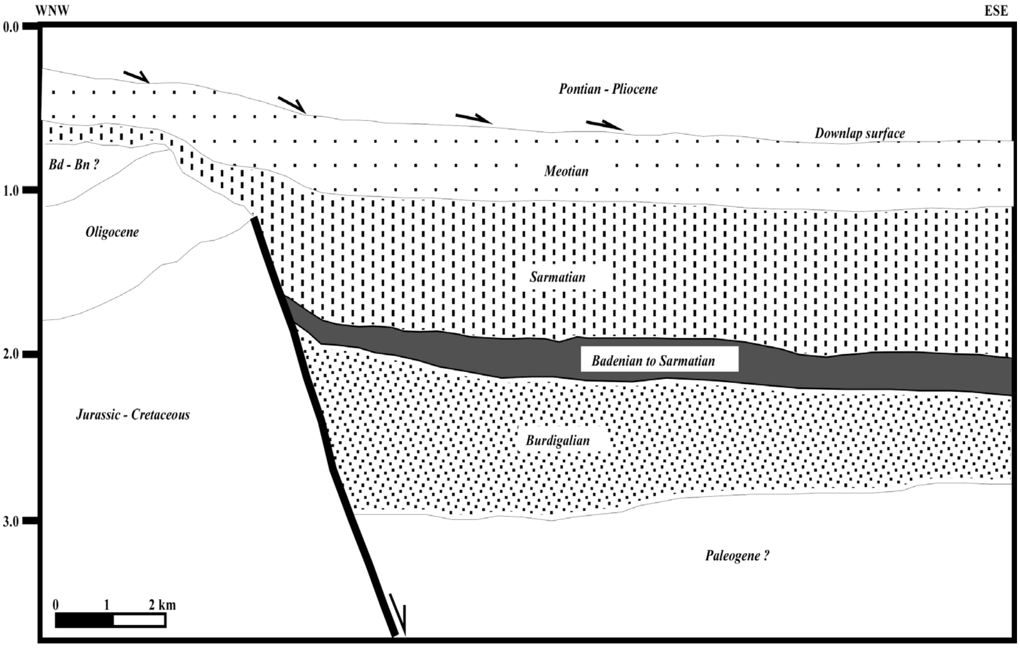


Figure 3.7 *Geological interpretation of a seismic line from the NW part of Moesia covering the westernmost fault-bounding extensional basin (modified after Răbăgia and Maţenco, 1999, their Figure 9). Note the major offset along the normal fault. The onset of extension is Early Miocene (Burdigalian), but Late Miocene (Sarmatian) reactivations are also indicated by the thickening of this sequence in the hanging-wall. The Meotian/Pontian boundary represents a downlap surface onto which large deltaic bodies prograded in a deepening basin (with a progressive reduction in the depositional energy from the Pontian onwards). Vertical scale is two-way travel time.*

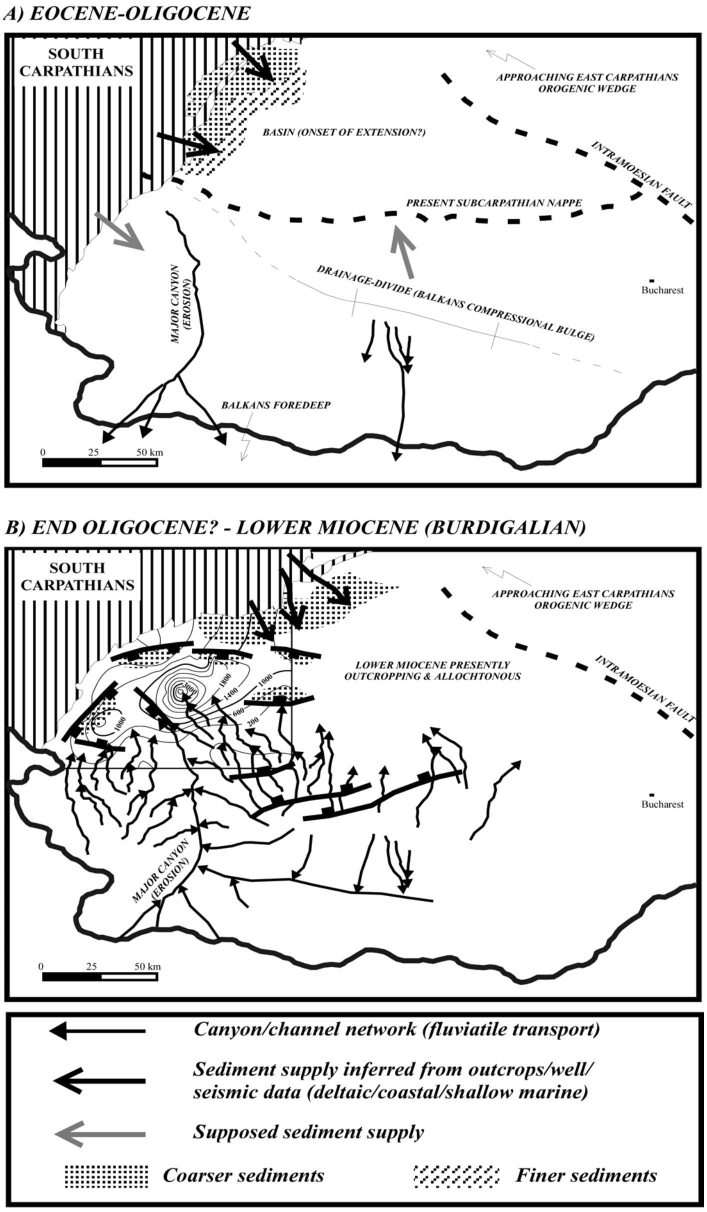


Figure 3.8 (**A**) *Eocene and Oligocene fluviatile networks derived from Figure 3.2A. The South-Carpathians are in their restored position as proposed by Schmid et al. (1998) and Maţenco and Schmid (1999). Direction of the sediment supply from the South-Carpathians (presently N to S) is based on the field studies of Jipa (1980, 1984).*
(**B)** *Lower Miocene (Burdigalian) fluviatile network derived from Figure 3.2A and its relationship with the depocentres of the (present) Getic Depression/northwestern Moesia. Contours show the Burdigalian subsidence according to Maţenco et al. (2003). Also shown is the Burdigalian fault system (from Figure 3.2A). The South-Carpathians are shown in the restored position according to Maţenco and Schmid (1999). The sediment supply directions from the South Carpathians nappe pile are inferred from the grain lithology and size observed in outcrops of the coarse Lower Miocene deposits.*

3.3.3. Age of erosion and driving mechanisms

The erosional period is obviously constrained by the age of the youngest sediments below the unconformity (Eocene on the southern Moesian margin) and the oldest sediments above it (Middle Miocene–Badenian). Paraschiv (1979b, 1997) considered that Moesia was a continental area during the Paleogene-Middle Miocene, the paleo-valleys flow direction being always N-wards. No inferences about mechanisms underlying the uplift required to produce such erosion were derived. These interpretations have not taken into account the changing morphological configuration, which resulted from the flexural tilting recorded by Moesia during the diacronous emplacement of the Balkans and Carpathians orogenic wedges on its margins (e.g. Fig. 3.1B).

Because the stratigraphic truncation increases N-wards (Figs. 3.5A and B), a Middle Eocene age is proposed for the onset of erosion, coeval with the emplacement of the Balkans thrust belt in the south (Doglioni et al., 1996), rivers having originally flowed S-wards (Fig. 3.8A) as still locally shown by the valley shapes (Fig. 3.2A). Accordingly, the uplift of the central-northern parts of Moesia could indicate its strong mechanical coupling with the orogenic wedge (sensu Ziegler et al., 2002) during the latest stage of the Balkans orogeny, which allowed that the compressional stress to be transmitted from the collision zone into the foreland accentuating a potential flexural fore-bulge.

During the Burdigalian, the flow direction of this drainage system was inversed with rivers now flowing N-wards, as demonstrated by the good correlation between the paleo-drainage network and the Lower Miocene subsidence map of the Getic Depression (Fig. 3.8B). This change may have started during the Oligocene, but no correlation can be made at this stage due to the relatively poor knowledge of Oligocene deposits, which are deeply buried below the present Getic Depression. The factors responsible for this change in the flow direction could be the (over)filling of the Balkans foredeep basin by the end of the Paleogene, the opening of Burdigalian extensional basin to the north and the post-orogenic uplift of Balkans-Srednogorie orogen. Overall, the available accommodation space was "transferred" from south to north. Continuation of incision during the Burdigalian was probably promoted by uplift of rift shoulder in the same area as the earlier, compression-related bulge (Figs. 3.8A, B and 3.9). The amount of uplift decreased from west to east, which correlates with the magnitude of the Early Miocene extension that followed the same direction (Răbăgia and Maţenco, 1999).

The Eocene ~N-S orientation of the drainage system, particularly of the major canyon, remained constant during the entire erosional period. This commonly happens if the uplift rate is lower than the incision rate of the drainage system, so that the river entrenches (Posamentier, 2001). The reversal of the flow direction from S-wards to N-wards along the major canyon pathway represents possibly the moment when the E-W oriented main tributaries (e.g. Fig. 3.8B) were formed by capturing some of the former S-wards flowing rivers.

Other mechanism may have enhanced the uplift of western Moesia, thus contributing to the observed amount of erosion. The Eocene core-complex formation in the South Carpathians (Maţenco and Schmid, 1999; Fügenschuh and Schmid, submitted) should induce unloading in the orogen and consequently, promote flexural rebound of the foreland. However, numerical modeling approaches are needed to validate the proposed processes responsible for the uplift and incision of western Moesia.

Although it is unusual to find such large canyons in foreland settings, the deep erosion in Moesia proves that apparently negligible vertical movements can induce large-scale features when suitably juxtaposed (Fig. 3.9), with significant impact on the basin evolution.

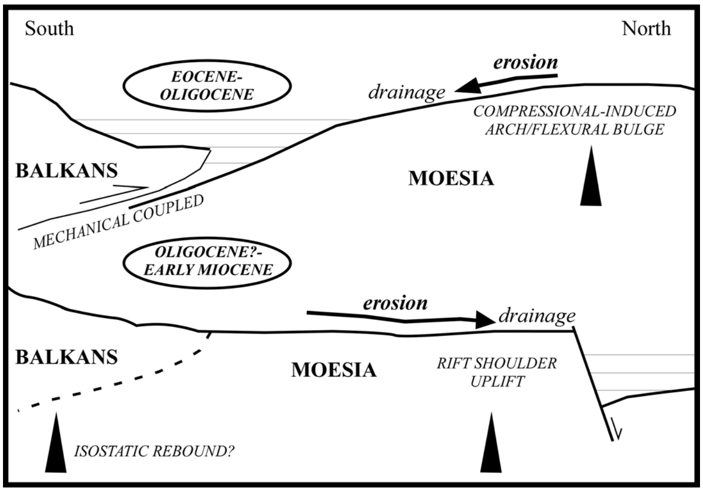


Figure 3.9 *Proposed mechanisms for uplift-driven erosion (not to scale). Compressional-induced bulge uplift is followed by rift shoulder uplift in the same region.*

3.3.4. Analogies with other regions

Deep erosional features resulting from interplay of different tectonic processes were described in other regions as, for instance, the foreland basin (Colville basin) facing the Brooks Range, Alaska (details in Cole et al., 1997) and western margin of Alès basin (Cévennes escarpment) belonging to the Alpine foreland basin of SE France (details in Roure et al., 1992; Sanchis and Séranne, 2000). In the former case, the E-W trending Colville basin is overthrust from the south by the Brooks Range orogen during Barremian-Aptian times. Younger contractions resulting in farther N-wards advancement of the thrust nappes are also documented. Significant erosion (in order of hundreds of metres) occurred to the northern end of the Colville basin, along the Barrow Arch. The onset of erosion pre-dated the Barremian-Aptian thrusting and was related to a Hauterivian extensional event taking place to the north of the Colville basin, which led to the opening of Canada basin and development of a break-up unconformity. Thus, the first mechanism responsible for the erosion was the extension-induced uplift of the rift shoulder (Barrow Arch). Further erosion of the northern margin of the Colville basin was promoted by flexural uplift in response of Barremian-Aptian orogenic loading and Early Tertiary renewed thrusting.

The latter example points out to the erosion occurring on the W-NW margin (part of the Massif Central) of the Alès basin, which is an Oligocene extensional basin superimposed on the Liassic rift structure of the Tethyan margin, later involved in Neogene compressional deformation. In the footwall of the Cévennes fault, up to 2 km of sedimentary section was removed by erosion such that the Hercynian basement largelly crops out in the Massif Central. The erosion was promoted by uplift of the rift shoulder during the Oligocene extension. Further continuation of the erosion was promoted by the Neogene contractions in the Western Alps through the uplift of a flexural fore-bulge and/or an intra-plate compressional stress-induced arch (sensu Ziegler et al., 2002).

3.4. REGIONAL SUBSIDENCE STAGE
3.4.1. Onset of deposition

Following the erosional stage, continental and later on marine sedimentation started in the Middle Miocene - Badenian (e.g. Paraschiv, 1979b; Chişcan et al., 1980). The Badenian deposits filled the former incised valleys coeval with the onset of the first contractional moment in the South-Carpathians foredeep, where NW-SE oriented thrusts of this age are documented particularly in its western part (Maţenco et al., 1997b; Răbăgia and Maţenco, 1999). The geological cross-section through the Getic Depression, published by the previously mentioned authors, show that Badenian sediments are thin, of the order of few tens up to a few hundreds of metres, with varying thicknesses. These sediments were mostly deposited in piggyback basins, except in the NW-most part of the Getic Depression where subsidence seems to have continued without any contractional deformation (Fig. 3.7).

The onset of continental sedimentation in most of western Moesia seems to have resulted from a decrease in the equilibrium profile of rivers in response to contraction-driven shallowing downstream and subsidence upstream. How much of this can be related to the inception of thrusting or to thermal relaxation of the lithosphere following Burdigalian extension, is impossible to determine at this stage. To derive this, one should look in the Paleogene basin in terms of sediment thickness and subsidence mechanisms. Consequently, modeling of the (known) Burdigalian extension together with the Paleogene tectonics will provide insights into the magnitude of the post-rift subsidence.

3.4.2. Filling the foredeep basin

The next stage in the basin filling history occurred in a typical foredeep setting as proposed by most authors (e.g. Paraschiv, 1979b, Săndulescu, 1984; Ionescu, 1994; Ionesi, 1994; Dicea, 1995; Popescu, 1995). Interpreted seismic lines apparently support this hypothesis (e.g. Figs. 3.5A, B and 3.6). However, continuation of the subsidence long after the final thrusting has received only recently attention (Maţenco et al., 2003; Bertotti et al., 2003). Consequently, a distinction is made in the following between the basin filling pattern/structure (1) during and, (2) post-dating the emplacement of Subcarpathian nappe (Late Miocene and Latest Miocene-to-Quaternary, respectively). The kinematics is also discussed as the structural features identified in this chapter differ from presently those consistent with accepted models of a simply flexure-related foredeep (e.g. Săndulescu, 1984).

3.4.2.1. Syn-thrusting Moesian basin

The thickness of the Late Miocene (Sarmatian) sedimentary sequence in western Moesia is laterally variable. The thickest sediments were deposited in the central part (Fig. 3.5A) whereas to the east thicknesses gradually decrease (Figs. 3.5B and 3.6). The basin shape in the central-eastern part of Moesia displays a N-thickening wedge, which can be genetically related to the emplacement of the Subcarpathian nappe. Sedimentation in this foredeep was contemporaneously with uplift of the inner region of the South-Carpathians (Fig. 3.10), as suggested by an angular unconformity at the base of Middle Sarmatian large coastal conglomeratic fans resting on top of Lower Sarmatian fine detritic sediments, which crop out at the contact between the nappe pile and sedimentary cover of the Getic Depression (inset in Figure 3.10), as well as by FT studies (Sanders et al., 1999). Sediments were also supplied from the south, mostly by a river network until the Late Sarmatian (Paraschiv, 1979b and references therein). The upstream area of this river network was possibly represented by the

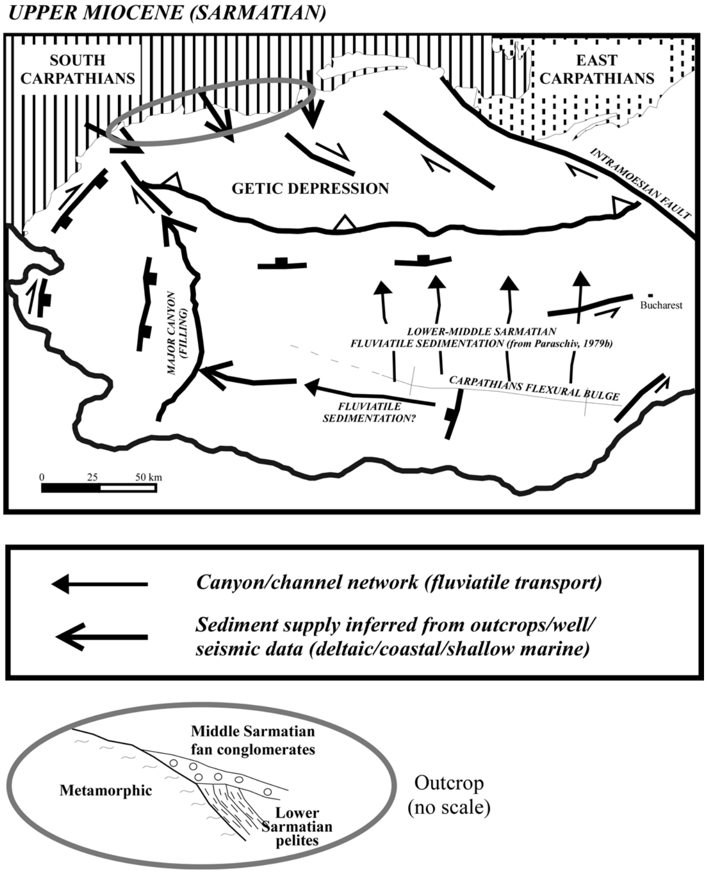


Figure 3.10 *Late Miocene (Sarmatian) pattern of sediment supply and active structures (simplified). At the contact with the internal nappe pile, large outcropping coastal conglomeratic fans (indicated by fauna) prograded towards the S and SE, coeval with the Middle Sarmatian emplacement of the Subcarpathian nappe (e.g. Săndulescu, 1988; Dicea, 1995). Uplift of the South-Carpathians can be inferred from this pattern, which is reinforced by the results of the FT analysis (cf. Sanders, 1998; Sanders et al., 1999).*

flexural bulge formed in response to the emplacement of the Subcarpathian nappe (Fig. 3.10). In contrast to the Paleogene-Burdigalian river erosion, no significant incision occurred during the Sarmatian, indicating that continental sedimentation kept pace with tectonic subsidence.

In the westernmost part of Moesia, basin subsidence was related to normal faulting rather than to flexural loading (e.g. Fig. 3.7). E-dipping normal faults with tens up to few hundreds of metres of offset were formed to the south of Burdigalian ones (Fig. 3.2A, see also Matreşu and Răbăgia, 2003). Also, the normal faults along the western margin of the major canyon may have been formed in Sarmatian times (Fig. 3.3B). At a regional scale, thrusting of the Subcarpathian nappe and NW-SE strike-slip faulting in the Getic Depression (e.g. Fig. 3.10) are interpreted to be the consequence of the general dextral movement between the intra-Carpathian units in the north and Moesia to the south (Maţenco et al., 1997b). Thrusting was progressively transferred to dextral strike-slip from east to west (also see Maţenco et al., 1997b). Minor flexure-related normal faults were formed parallel with the Subcarpathian thrust (e.g. Fig. 3.2A). Strike-slip movements may also be interpreted along faults in the central and eastern parts of western Moesia (Figs. 3.2A, 3.5B and 3.6). Sinistral strike-slip faults can be interpreted as antithetic Riedels to the coeval large-scale movement of the dextral Intramoesian fault (see also next chapter and Tărăpoancă, 1996). The sinistral sense is proposed because these faults offset the Permo-Triassic rift structures (Răbăgia and Tărăpoancă, 1999).

The kinematics of normal faulting (Figs. 3.2A, 3.3B and 3.7) in the westernmost part of Moesia is less easy to explain. Because the Sarmatian period represents the time of E-wards transport of the intra-Carpathians units (e.g. Mațenco et al., 1997b) and based on a few seismic images, dextral transcurent movements were proposed (Fig. 3.10; Mațenco et al., 1997b; Răbăgia and Mațenco, 1999; Matreşu and Răbăgia, 2003). However, we cannot be sure whether the dextral transtension in the westernmost part of Moesia is compatible with the dextral transpression in the Getic Depression produced by the E-ward movement of the Carpathians units.

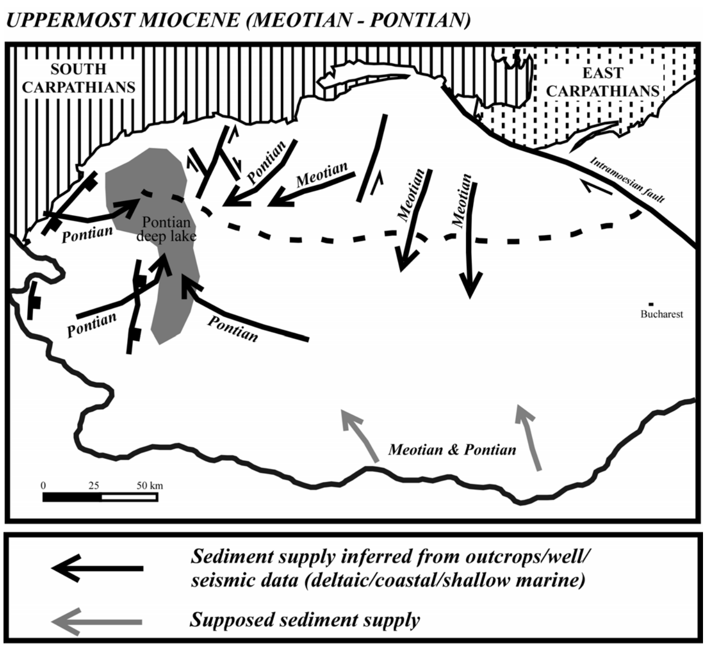


Figure 3.11 *Latest Miocene pattern of sediment supply and active structures (structures within Getic Depression are simplified after Răbăgia and Mațenco, 1999).*

3.4.2.2. Post-thrusting Moesian basin

Following the emplacement of the Subcarpathian nappe, subsidence of Moesia continued through the Latest Miocene (Meotian-to-Pontian) and Pliocene-Quaternary. During the first part of the Latest Miocene (Meotian), most sediments were supplied to the basin from the north (Figs. 3.5B and 3.11), indicating active erosion of the South-Carpathians/Getic Depression. In Moesia, the Meotian sequence has a relatively constant thickness (Figs. 3.3 – 3.7) that suggests that uniform subsidence was taking place. Maximum subsidence is recorded in the NW-most part and is mimic the previous Burdigalian depocentres (Bertotti et al., 2003) as suggested by directions of the sediment supply (Fig. 3.11). This subsidence is not linked to significant tectonic activity, because only minor deformation is observed in the Getic Depression (e.g. Răbăgia and Maţenco, 1999) and only small-scale normal faulting was active in the westernmost Moesia (e.g. Fig. 3.7).

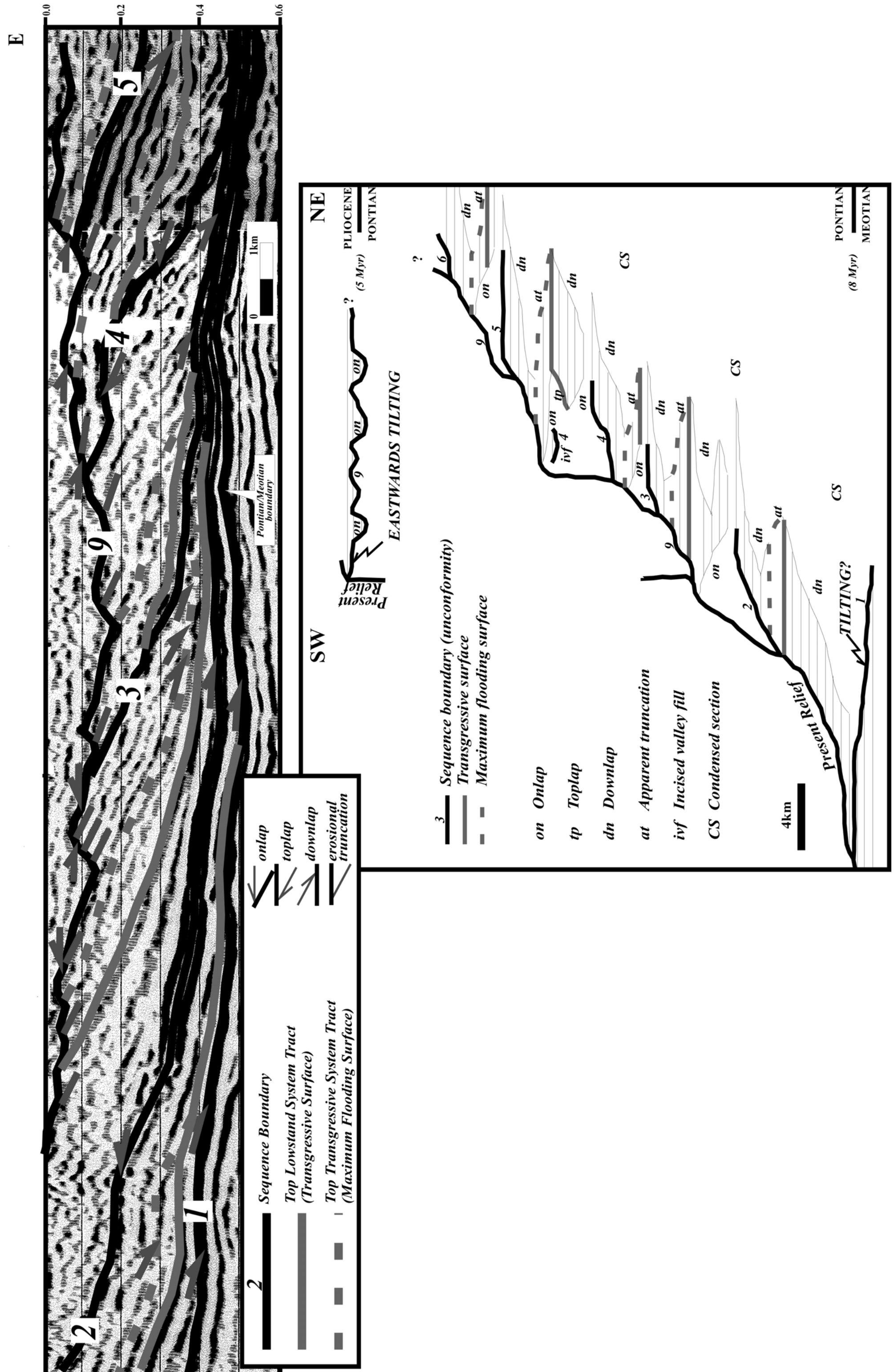
An important change in the sedimentation pattern took place at the Meotian/Pontian boundary in the central-western part of Moesia. Here, large deltas with dominant prograding seismic facies started to develop (Figs. 3.3A and B, 3.4 and 3.7). These deltas supplied the sediments mainly to the NW-most part of Moesia (Fig. 3.11), which continued to function as the most subsiding area (Bertotti et al., 2003). The foresets configuration of the Pontian sequence indicates a progradation in lacustrine environments (according to Saulea et al., 1969, Fig. 2.7) as deep as ~300 m (note that the difference between topsets and corresponding bottomsets is ~300-350 ms in Figures 3.3 and 3.4). The progradation appears to overlie the area of the previous major canyon (Fig. 3.3B; also compare Figs. 3.11 with 3.8). We relate this increase in accommodation space to the downward movement of the hanging-wall of the normal fault system from the western flank of the major canyon (Fig. 3.3B).

The location of the Pontian source areas shown in Figure 3.11 contrasts with previous studies that considered the NNW-to-NW part of the South-Carpathians as providing most of the sediments (Jipa, 1997). Comparison of the Pontian supply pattern with the Meotian one suggests that the source area moved from the central-eastern South-Carpathians/Getic Depression along the arc to the western South Carpathians/Balkans and maybe southernmost Moesia (Fig. 3.11). This may reflect lateral changes in uplifted segments of the orogen or, the breaking of the Danube through the South-Carpathians. However, as in the Meotian, Pontian subsidence occurred without significant tectonic deformation in the orogen according to Săndulescu (1988), Ratschbacher et al. (1993), Maţenco et al. (1997b), Hippolyte et al. (1999), Răbăgia and Maţenco (1999) or Maţenco et al. (2003).

A detailed interpretation of the thick prograding bodies in the westernmost Moesia is shown in Figure 3.12. Sequential analysis of the contact zone between the basin (Moesia) and the orogen nappe pile provides constraints both on the last stages of basin filling and on the onset of fluviatile sedimentation related to the development of the paleo-Danube. Several sequences can be related to variations in the base level, which are probably tectonic rather than eustatic-driven. Generally, the lowstand wedge thickness decreases from the older to younger sequences (from west to the east) whereas the highstand system tracts thicken. Also, the seismic facies show an oblique-parallel configuration in the west changing progressively to

Figure 3.12 (next page, up) *Detailed interpretation of the shallow part of the seismic line shown in Figure 3.4 (westernmost part of Moesia, next to the contact with the outcropping South-Carpathians nappe pile). Several sequences can be identified within the Pontian. Note that the base Pontian represents a downlap surface. The two basal Pontian sequences are cropping out at the westernmost part of the line. A well-expressed erosional unconformity corresponds to the Pliocene/Miocene (Pontian) boundary. This widely incised valley witnesses the onset of the paleo-Danube River flow to the east of the South-Carpathians.*

Figure 3.13 (next page, down) *Chronostratigraphic chart of Pontian-to-Pliocene clinoform sequences in westernmost Moesia. Chart direction is shown in Figure 3.2A. The sequential analysis is based on the interpretation of a seismic survey having a length of a few hundred kilometres. Nine unconformities (type 1 sequence boundaries) are identified. Pliocene erosion affected deeper levels to the west, indicating E-wards tilting. Unconformity 9 is onlapped by fluviatile deposits of the paleo-Danube River, which subsequently migrated towards its present position.*


W
E
SW
NE
Sequence Boundary
Top Lowstand System Tract (Transgressive Surface)
Top Transgressive System Tract (Maximum Flooding Surface)
onlap
toplap
downlap
erosional truncation
Pontian/Meotian boundary
1km
Sequence boundary (unconformity)
Transgressive surface
Maximum flooding surface
on Onlap
tp Toplap
dn Downlap
at Apparent truncation
ivf Incised valley fill
CS Condensed section
4km
Present Relief
EASTWARDS TILTING
TILTING?
PLIOCENE
PONTIAN
MEOTIAN
(5 Myr)
(8 Myr)

oblique-tangential and oblique-sigmoid to the east. This pattern clearly indicates a river system prograding into a lake. A regional erosional unconformity truncates the Pontian deltaic sediments in the westernmost part of Moesia (Fig. 3.12). This erosional stage is related to an E-wards tilting than to shoreline regression. This conclusion is illustrated by the chronostratigraphic chart (Fig. 3.13), which shows that the erosion reaches to the lowstand system tracts of the older sequences in the western area. To the east, the erosion removes only parts of the former transgressive and highstand system tracts. It is difficult to provide an estimate of the missing section to the west; nevertheless, it appears that before this erosional event, the basin had extended farther to the west overlying areas from the South-Carpathians that have present elevations as high as ~400 - 500 m.

Starting in the Pliocene, the Moesian basin has been filled by continuously E-wards progradation in which the fluviatile facies has progressively replaced deltaic and lacustrine environments (e.g. Jipa, 1997). In the central-eastern part of the basin, the Pliocene-Quaternary sequence thickens towards the front of the Subcarpathian nappe (Figs. 3.5A, B and 3.6). Overall, it thickens from west to east, in contrast with the Sarmatian trend. Figure 3.6 shows that regional N-ward tilting of Moesia was apparently achieved mostly during Pliocene-Quaternary rather than during Sarmatian (i.e. syn-thrusting). The last stage of subsidence in Moesia was coeval with the uplift in the internal South-Carpathians and minor thrusting in the eastern part of Getic Depression (Maţenco et al., 1997b). This correlates with the N-wards tilting of Moesia increasing to the east. At this stage, most of the active tectonics is related to reactivated ENE-WSW oriented fault systems affecting the central-eastern part of basin (e.g. Fig. 3.6), interpreted as strike-slip faults that offset sinistrally the Permo-Triassic rift structures (Răbăgia and Tărăpoancă, 1999). Some normal faults reactivations are also documented in the westernmost part of Moesia (Figs. 3.2A and 3.3B; also Matreşu and Răbăgia, 2003).

3.4.3. Syn- and post-thrusting subsidence: driving mechanisms

One possible explanation for the persistence of subsidence of foreland basins after thrusting had ceased may be related to pre-contractional crustal extension (Deségaulx et al., 1991; also see Chapter 2). Depending on the interplay between the amount of extension and the age of thrust loading, thermal cooling subsidence can be “added” to the flexural component. This could be one of the mechanisms responsible for part of the overall subsidence in western Moesia, as extension occurred at least during the Burdigalian, prior to the Middle Sarmatian emplacement of the Subcarpathian nappe. At least, the Badenian-Early Sarmatian subsidence could be related to such a thermal cooling process, because the orogenic loading was minor at that time and probably not enough to account for the change from high relief to marine sedimentation. This mechanism might also have contributed to the Late Miocene subsidence of the western Moesia. The deepening of the NW-most part of Moesia that is inferred from the seismic facies (e.g. Fig. 3.3) for the Latest Miocene (especially for the Pontian) can partly result from the tectonic activity of the ~N-S trending normal faults. How much the compaction of sediments contributed to the overall subsidence is also unknown. Discrimation between the compaction and tectonic components to the total subsidence requires a detailed sequential analysis, integration of lithological and paleontological data to derive paleo-water depths, as well as a flexural rather than an Airy backstripping approach.

The Pliocene-Quaternary subsidence appears as problematic since post-rift cooling is not compatible with the uplift documented in the orogen (Maţenco et al., 1997b). Moreover, no significant orogenic loading (thrusting) occurred coeval with this subsidence. Hence, another mechanism is required to explain the last subsidence stage of western Moesia. Because in many cases the lithosphere responds to compression by folding, with relatively large wavelength-spaced uplift and subsidence (Cloetingh, 1988; Cloetingh et al., 1999), a buckling-related subsidence has been recently proposed for Moesian region (Bertotti et al., 2003). This mechanism is compatible with the Pliocene-Quaternary vertical movements in the Carpathians realm (see also Chapter 6), and with the Pliocene-Recent NNW-SSE oriented compressive stress (e.g. Mațenco et al., 1997b; Bada et al., 1998; Hippolyte et al., 1999). It appears that the buckling-related subsidence of the Moesian region was contemporaneously with that of the centre of the Pannonian basin, for which Horváth and Cloetingh (1996) proposed a similar origin. It is restated that these mechanisms proposed for the evolution of Moesia, although plausible, should be further validated through modeling studies. If the syn- and post-thrusting subsidence in western Moesia were really the sum of flexural loading, post-rift cooling and buckling, the Sarmatian-Pliocene normal faulting in the westernmost part would accommodate the differential vertical movements between the orogen and the foreland.

3.5. CONCLUSIONS

The western Moesia had a complex Tertiary evolution, which can be basically divided into two stages with contrasting patterns of vertical movements: Paleogene-Early Miocene and post-Early Miocene. During the first stage, most of western Moesia (except its northern and southern margins) was uplifted and deeply eroded by an extensive river network. Uplift and river incision started during the final thrusting of the Balkans over the southern Moesian margin in the Middle Eocene and continued during the movement and rotation of the South-Carpathians units towards their present position. Erosion was triggered by uplift of the area which is presently oriented ~E-W in the middle of the Moesian block due to strong mechanical coupling between the Balkans orogenic wedge and its foreland that accentuated a potential flexural fore-bulge. Accordingly, the compressional stresses build-up upon collision could be transmitted to the foreland (Ziegler et al., 2002) leading to arch development. Later on, erosion was promoted by Early Miocene extension/transtension, which affected the northern margin of Moesia (Getic Depression) including the Carpathians nappe pile (e.g. Răbăgia and Mațenco, 1999). Most probably, uplift of this southern rift shoulder affected more or less the same region as the previous compression-related arch.

The second evolutionary stage started in the Middle Miocene (Badenian) when sedimentation commenced on the previous eroded areas coeval with the onset of contraction in the deformed foredeep of the Getic Depression. The Late Miocene (Sarmatian) pattern of vertical movements relates mostly to the emplacement of the Subcarpathian nappe on the northern parts of Moesia and partly to the normal faulting along its westernmost part. During the Latest Miocene, subsidence of western Moesia was accompanied by erosion of the inner foredeep in the absence of significant contractions in the orogen. Pliocene-Quaternary subsidence of the foreland was coeval with large-scale uplift of the inner South-Carpathians and minor contractions in the inner part of the foredeep (e.g. Mațenco et al., 1997b). Three mechanisms (post-rift cooling, flexure and buckling) are thought to have possibly contributed to this pattern of vertical movements, which partly post-date thrusting. Thermal changes originating in the Burdigalian rifting are postulated to be of primarily importance in explaining subsidence prior to the Sarmatian thrusting whereas lithospheric buckling seems to best account for the Pliocene-Quaternary subsidence (Bertotti et al., 2003).

Overall, the interplay between different tectonic processes occurring in a highly curved plate boundary setting is probably the cause of large-scale opposing vertical movements in foreland basins. Feedback between each tectonic process and the induced rheological changes in the lithosphere essentially controls the pattern of successive deformations.

CHAPTER 4

THE EAST-CARPATHIANS FOREDEEP/FORELAND. THE ARCHITECTURE AND SUBSIDENCE OF THE FOCŞANI DEPRESSION

4.1. INTRODUCTION

The East-Carpathians foredeep records variable amounts of subsidence, with the highest values being associated with changes in the orogenic strike, i.e. in the SE-Carpathians Bend region (Focşani Depression). Significant spatial and temporal variations are documented also along the East-Carpathians foreland basin and these can only partly be related to the typical orogenic deformations occurring within the thrust-and-fold belt. Specifically, parts of the foreland basin are characterized by anomalously large subsidence following the main contractional event in the orogen. The orogenic deformations are also variable, from relatively simple in front of the central-northern East-Carpathians to highly complex in the SE-Carpathians Bend foreland, in terms of timing, geometry and stress regime. Previous studies of the East-Carpathians foreland (e.g. Gavăt et al., 1969; Paraschiv, 1979a; Săndulescu, 1984; Săndulescu and Visarion, 1988; Visarion et al., 1988) have presented variable structural pattern (Fig. 4.1), but paid little attention to timing of deformation or structural style.

The Focşani Depression is developed next to the SE-Carpathians Bend, where the NNW-SSE trending East-Carpathians meet the E-W striking South-Carpathians. Hosting >10 km thick sediments (Gavăt et al., 1969; Dumitrescu and Săndulescu, 1970; Rădulescu et al., 1976; Dicea, 1995), the Focşani Depression developed in Neogene to Quaternary times, i.e. during and, more importantly, following the main contractional stages of the Carpathians. The Carpathian Bend Zone and the Focşani Depression are of special interest because they correspond to the site of the youngest deformations in the Carpathians domain (Săndulescu, 1984; Hippolyte and Săndulescu, 1996) and to some of the most seismically active regions, namely the Vrancea zone with its cluster of seismicity at depths of 40–200 km (Oncescu, 1984; Oncescu et al, 1988). Whereas the structure of the central-northern East-Carpathians foredeep is relatively well known (e.g. Paraschiv, 1979a), the geometry of the Focşani Depression and especially its internal architecture are poorly constrained. To better understand the very great thickness of the sediments in this depression and its significant subsidence post-dating thrusting, a detailed knowledge of the architecture and development of the Focşani Depression is required. Previously proposed evolutionary models for the Carpathians (see Chapter 2) rely on inferences drawn from cross-sections through the Bend region and the Focşani Depression.

The purpose of this chapter is to provide new data on the East-Carpathians foredeep/foreland, particularly for the SE Bend zone and eastern Moesia. As long as the detailed foredeep architecture and development of the Focşani Depression have not been established, the link between Carpathians deformations, Focşani Depression subsidence and Vrancea seismicity cannot be understood. Consequently, the core of this chapter is the documentation of the 3D architecture and the subsidence evolution of the Focşani Depression and the Bend foreland. The different behaviour of the East-European/Scythian platform in the north and of Moesia in the south is highlighted, both in terms of vertical movements and

This chapter is mainly based on: Tărăpoancă, M., G. Bertotti, L. Maţenco, C. Dinu and S.A.P.L. Cloetingh, Architecture of the Focşani Depression: a 13 km deep basin in the Carpathians Bend zone (Romania); Tectonics, v. 22, p. 1074-1092; and

Tărăpoancă, M, C. Dinu and D. Ciulavu, Neogene kinematics of the northeastern sector of the Moesian platform (Romania); AAPG Bulletin, in press.

structural pattern. This pattern will be integrated in a regional scale in Chapter 6 in an attempt to correlate the vertical movements of the foreland with those determined for the orogen (Fig. 4.2; see also Bertotti et al., 2003).

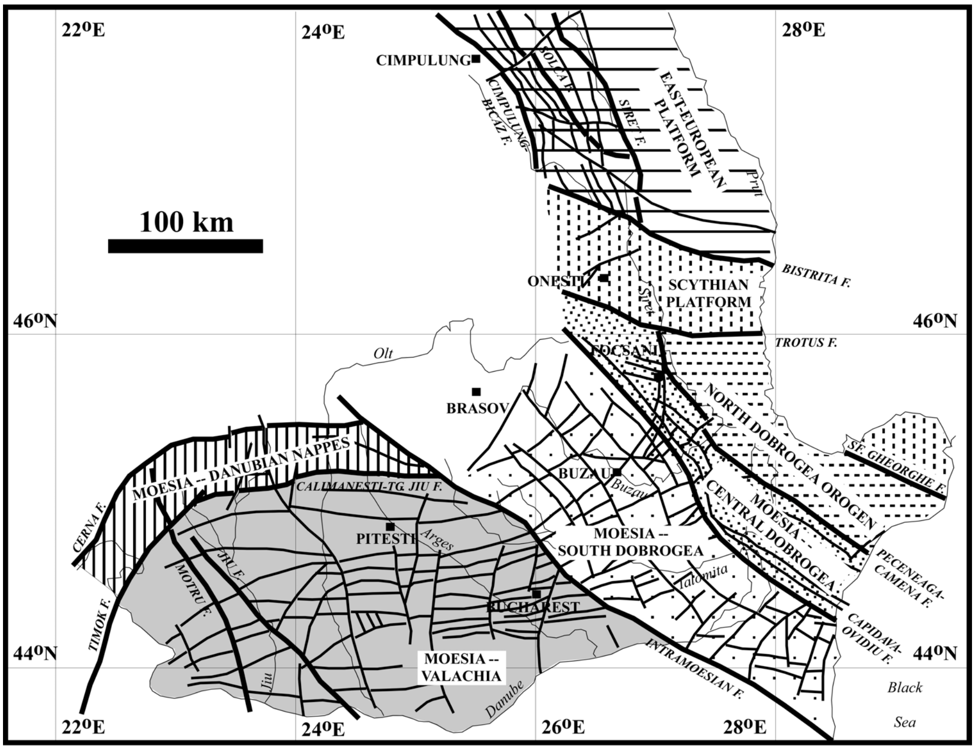


Figure 4.1 *Structural pattern of the Carpathians foreland (references in text).*

4.2. DATABASE

Over 1000 km of 2D seismic lines were used (Fig. 4.3) to establish the 3D geometry of the SE Carpathians Bend foreland and its internal architecture, as well as its Neogene-Quaternary tectonic evolution. More than 60 wells were used to calibrate (dating) seismic horizons and derive information on seismic velocities. These wells are located around the Focşani Depression, except for its western margin. No well has been drilled so far inside the Focşani Depression. Therefore, the velocity maps used for depth-conversion, contain interpolated values within this basin. Generally, for each sequence, the greatest velocity resulted within Focşani Depression due to the deepest burial and, hence, maximum compaction of sediments. The influence upon the velocity profile of overpressured zones that may be present in the thick sedimentary column was not considered.

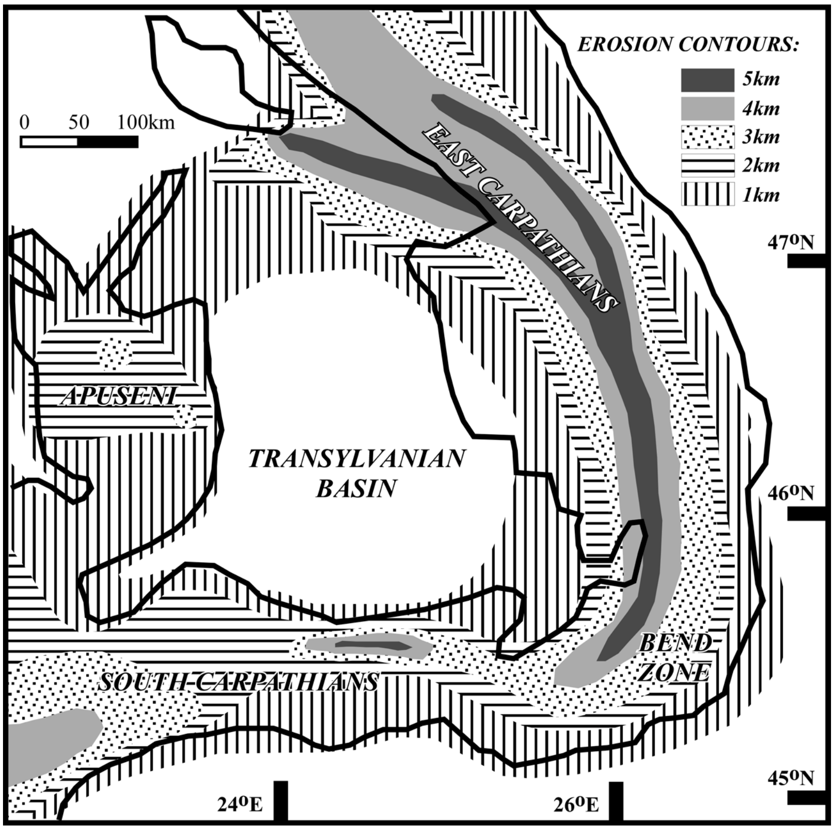
Five seismic horizons were interpreted and mapped: Base Tertiary, Top Badenian, Top Sarmatian, Top Meotian and Top Pontian (= Base Pliocene). The time-structure map of each horizon was depth-converted using average velocity maps instead of interval velocities. This approach is appropriate because not all sequences cover the entire studied region and because of the large dimension of the area. Finally, isopach maps were constructed by subtracting the depth-structural map of the top of the respective sequence from the map of its base. Total subsidence curves were extracted from these maps in several locations (shown as dots in Fig. 4.3).

We preferred to not construct tectonic subsidence curves because some of the required input data cannot be yet properly constrained in terms of petrophysical characteristics and paleo-bathymetries (especially in the Focşani basin where no well data exist) and because the backstripping method that is usually used (Airy) is not appropriate for a setting where the flexural mechanism appears to exert an important control. Specifically, a sequence stratigraphic approach is required in order to constrain well the vertical changes in lithology and paleobathymetry before removing the contribution of the sediments and water loading to the overall subsidence. However, a detailed sequence analysis is beyond the aims of this study.

In the SE-Carpathians foreland the western limit of the maps constructed corresponds to the Pliocene-Pleistocene Caşin-Bisoca reverse fault, which is conventionally taken as the western border of the Focşani Depression (Dicea, 1995). The Late Miocene (Sarmatian-Pontian) deposits in the vicinity and to the west of this fault are tilted towards the east with steep to vertical dips and are partly eroded (Dumitrescu et al., 1970). This not only makes it difficult to provide accurate thickness estimates, but also leaves open the question of the real western termination of the basin. Therefore, this conventional limit is valid just for the present time as the actual geometry of the western margin indicates that, geographically, the Late Miocene Focşani basin extended farther to the west over the Subcarpathian nappe. Instead, geometries at the eastern margin of the Focşani Depression are well constrained thanks to the shallow basin depth and numerous industrial wells.

The geometry of the central-northern East-Carpathians foreland (in the area of the East-European platform) is modified from Paraschiv (1979a, 1989). The structure is simpler than in the Moesian domain, with a shallow typical wedge-type foredeep basin. The foredeep stage on the East-European platform spans only the Sarmatian period. For the southern part of Moesia, no thickness map is provided at this stage, because the density of reliable seismic surveys is too little to derive an accurate 3D geometry, although it is sufficiently good to identify the structural pattern.

Figure 4.2 (next page) *Map showing amount of sedimentary column eroded from the Carpathians, based of FT analysis (Sanders, 1998; Sanders et al., 1999).*



4.3. PRESENT-DAY GEOMETRY OF THE EAST-CARPATHIANS FORELAND AND STRUCTURAL RELATIONSHIP WITH THE OROGEN

The base Tertiary map in the East-Carpathians foreland is shown in Figure 4.4A. The structural grain of the Romanian East-Carpathians foreland has generally a distinct NNW-SSE trend. Relative to the orogen, the structures of the East-European platform parallel the thrust belt up to the Trotuş fault, becoming oblique more to the south. The latter area represents not only the junction of four different lithospheric blocks but is also the site of the greatest foredeep subsidence (i.e. Focşani Depression; Fig. 4.4B).

Coming along strike from the Polish East-Carpathians, where the foredeep has depths of up to 1.5-1.7 km (Krzywiec, 2001), the depth to the base Tertiary unconformity decreases to ~1km in the northern part of Romania near the main thrust front. Farther southwards, the sediment thickness increases from NE to the SW and then S near the thrust front (Fig. 4.4A; Paraschiv, 1979a, 1989; Dicea, 1995). Sediments shallow eastwards and crop out near the Prut River. A fault system with vertical offsets in order of several hundreds of meters parallels the central-northern East-Carpathians front.

Two kilometres increase in sediment thickness occurs along the WNW-ENE trending Trotuş fault (Fig. 4.4A). In contrast to published maps of this area (Fig. 4.1; e.g. Săndulescu, 1984; Săndulescu and Visarion, 1988), we interpret the “Tertiary” Trotuş fault as ending towards the east and its position is located 25-30 km farther to the north than previously considered.

Sediment thicknesses reach up to 13 km in the NNW-SSE oriented Focşani Depression, slightly north of the Bend Zone and south of the Trotuş fault (Fig. 4.4A). The geometry of the Focşani Depression is constrained by several transversal seismic lines, as the one shown in Figure 4.4B. Whereas the shallower markers can be accurately interpreted due to correlations with subsurface information on the eastern margin (North Dobrogea promontory) and outcrop observations on the western one, the deepest horizons, particularly the base Tertiary, are sometimes interpreted by analogy with the overall shape of the basin. This approach applies to the centre of the basin, where the Badenian sequence partly lies below 5-5.5 s, which represents the bottom of the time-scale of the seismic lines.

The thickness of the sedimentary infill of the Focşani Depression gradually decreases both towards the ENE and to the SSE (Fig. 4.4B). The western continuation of the depression is deformed and partly buried under Carpathians thrusts and is difficult to constrain. The transition from the Focşani Depression to the E and NE platform areas (i.e. North Dobrogea promontory) is affected by a NW-SE trending fault system that extends south of the Trotuş fault (Fig. 4.4A).

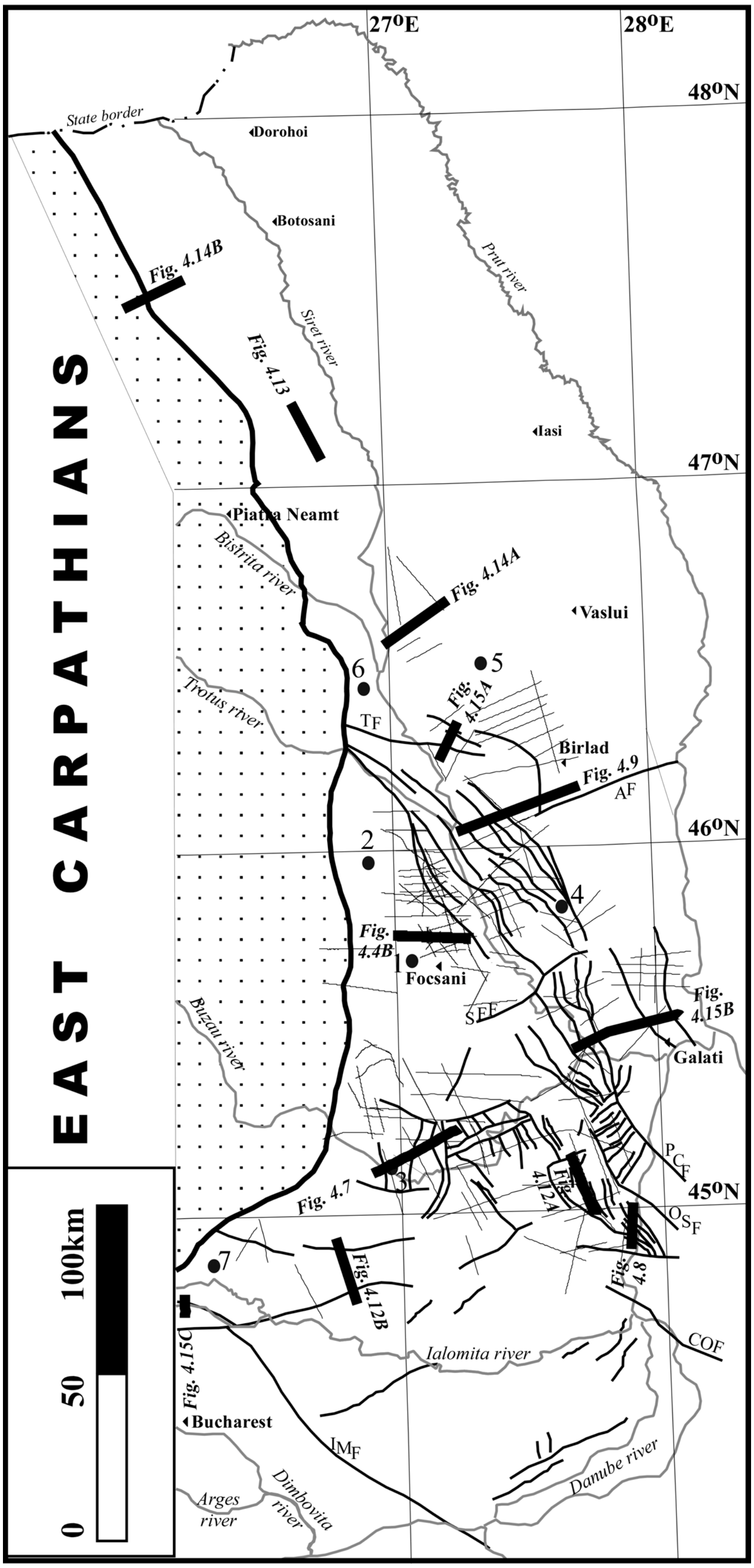
To the SE of the S-Focşani fault, the North Dobrogea promontory is uplifted by 1 km or more with respect to Moesian platform along the Peceneaga-Camena fault. The southern margin of the Focşani Depression is more gradual than the eastern one. N-S to NW-SE trending sub-basins are observed south of the Focşani Depression (Fig. 4.4A) and will be described in more detail below. The western portions of the southern border are affected by the Intramoesian fault which displays a vertical offset of >1km in the NW decreasing towards the foreland.

The contact between the orogen and the undeformed foredeep is represented by the frontal thrust of the Subcarpathian nappe (named Pericarpathian fault), which crops out north of the Trotuş fault and cuts Lower-Middle Sarmatian autochthonous sediments and is sealed in few locations by Upper Sarmatian deposits (examples in e.g. Săndulescu, 1984; Ştefănescu et al., 1988; Ionescu, 1994; Dicea, 1995; Maţenco and Bertotti, 2000).

Below the Subcarpathian nappe, the autochthonous sediments of the East-European platform are clearly recognized to depths of up to 3.6 km to the west (Fig. 4.4A; see also Paraschiv, 1979a). Between the Trotuş and Buzău rivers (Fig. 4.3), the front of the Subcarpathian nappe is buried beneath deposits younger than Middle Sarmatian. In this region, the contact between the frontal nappe zone and the Upper Sarmatian deposits is considered to be either stratigraphic (e.g. Săndulescu, 1984) or a backthrust defining a triangle zone (Maţenco and Bertotti, 2000; Fig. 2.3B).

Between Buzău valley and the Intramoesian fault (Fig. 4.3), the Subcarpathian nappe is interpreted as being covered by Upper Sarmatian deposits (e.g. Maţenco, 1997). According to Maţenco and Bertotti (2000), the minimum displacement of the Subcarpathian nappe above the undeformed foreland can be estimated at more than 15-25 km everywhere along the East-Carpathians.

Figure 4.3 (next page) *East-Carpathians foreland: location of the seismic survey (thin lines) used for interpretation and mapping (color maps shown in Figures 4.4, 4.5, 4.11, 4.16 A, B and C). Thick black lines show the structural pattern referred to in this chapter. Thickest black lines represent seismic sections used as examples with the corresponding figure number. Black numbered dots represent the location of subsidence curves shown in Figure 4.6.*



27ºE
28ºE
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47ºN
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45ºN
State border
Dorohoi
Botosani
Prut river
Siret river
Iasi
Piatra Neamt
Bistrita river
Vaslui
Trotus river
Birlad
Focsani
Galati
Buzau river
Ialomita river
Bucharest
Danube river
Arges river
Dimbovita river
EAST CARPATHIANS
Fig. 4.14B
Fig. 4.13
Fig. 4.14A
Fig. 4.15A
Fig. 4.9
Fig. 4.4B
Fig. 4.15B
Fig. 4.7
Fig. 4.12A
Fig. 4.8
Fig. 4.12B
Fig. 4.15C
TF
AF
SFF
PCF
OSF
COF
IMF
1
2
3
4
5
6
7
0
50
100km

4.4. ARCHITECTURE, ACTIVE STRUCTURES AND BASIN EVOLUTION OF THE FOCŞANI DEPRESSION AND THE EAST-CARPATHIANS FORELAND

4.4.1. Middle Miocene (Badenian)

4.4.1.1. Thickness and subsidence patterns

The Badenian marks the onset of subsidence in the East-Carpathians foreland including the Focşani Depression. Although no well penetrated the Badenian sequence in the Focşani Depression, the well data from the East-European/Scythian platform and the shallower parts of eastern Moesia/North Dobrogea promontory indicate a shallow water environment based on the evaporitic (anhydrite) layers interbedded in clastics and rare carbonates (Saulea et al., 1969; Paraschiv, 1979a). The general pattern is one of increasing thicknesses and subsidence towards the west (Fig. 4.5). The Focşani Depression represents the region with the greatest subsidence. Very high thicknesses up to >4 km and subsidence rates of >1 km/Myr (curves #1, 2 and 3 in Fig. 4.6) are recorded in its western parts. Isopachs have a distinct NNW-SSE trend in the east and a NE-SW trend in the south.

Although with much lower thicknesses than in the Focşani Depression, Badenian sediments are also found on the East-European/Scythian platform. Their presence demonstrates that subsidence affected large areas of the mechanically strong and competent East-European platform. North of the Trotuş fault and along the present Carpathians front, Badenian deposits are 0.2-0.4 km thick. These thicknesses decrease progressively to the NE to a few tens of meters. In the footwall of the thrust bounded by the Adjud fault, sediments reach thicknesses of 0.5-0.6 km (Fig. 4.5).

E-NE of the Focşani Depression (towards the North Dobrogea promontory), thickness decreases rapidly to values as small as 0.1-0.2 km. Subsidence rates are correspondingly very low and are typically 10 times lower than in the central domains (Fig. 4.6). Farther E-wards, the Badenian sequence pinches out.

To the SE of the Focşani Depression, fault-bounded basins control the distribution of Badenian deposits. Thicknesses range between 1.2 to 2.5 km in the northern fault-bounded basins and gradually decrease SE-wards (Fig. 4.5). Some of the extensional “rift” shoulders of these basins were emergent during the Badenian. Towards the SW, Badenian sediments are roughly 0.1-0.2 km thick near the Intramoesian fault and pinch out to the south.

4.4.1.2. Active structures

The most apparent structures active during Badenian times are N-S to NW-SE trending normal faults located mainly along the SE margin of the Focşani Depression (Fig. 4.5). These faults are steep and define N-S and NW-SE trending grabens, half grabens and horsts (Figs. 4.7 and 4.8). A narrow and deep graben marks the transition from N-S trending grabens in the west to those trending NW-SE (Figs. 4.5 and 4.7). The width and depth of grabens decrease towards SE. Extensional faulting propagated from NW to SE and the Badenian syn-rift sequence becomes younger in the same direction. The grabens are bounded by ENE-to-E trending transfer faults (Fig. 4.5). During the Middle-Late Badenian, mild inversion started in the westernmost extensional basin and lasted until the beginning of Sarmatian (Fig. 4.7). The Badenian/Sarmatian boundary is represented by an erosional unconformity of regional extent (Fig. 4.7).

A possible prolongation of these normal faults towards the NW can be demonstrated only for the eastern parts of the Focşani Depression where Badenian sediments are quite shallow. NW-SE trending, mainly SW-dipping normal faults follow, for instance, the margin of the depression north of the Buzău River (north of S-Focşani fault). Offsets are in the range of tens to a few hundreds of meters (Fig. 4.5). They extend to the NW beneath the present Carpathians structures. Areas to the NE of the normal faults formed part of their footwall and were partial emergent during the Badenian (note the narrow NW-trending area with no sediments in Figure 4.5).

On the East-European/Scythian platform, a NW-SE trending thrust fault system formed during the Badenian, which is coeval with the extension in the southern foreland (Figs. 4.5 and 4.9). This system is formed by three thrust faults, the longest one having a curved shape. The dip of the faults increases progressively to the NW where the offset decreases. Piggy-back and flexure-associated basins developed (Fig. 4.9) related to thrusting. The thrust fault system is bounded to the south by the Adjud fault (interpreted as sinistral strike-slip fault), which extends beyond the eastern border of studied area (Fig. 4.5). One can speculate that these thrust faults developed in a restraining stepover setting between the Adjud and Trotuş faults. This would imply sinistral displacements along the latter as well. However, no kinematic indicators have been found so far west of the present Carpathians front to attest to Badenian tectonic activity along the Trotuş fault (cf. Maţenco, 1997; Maţenco and Bertotti, 2000).

Northwards, no significant fault system was identified for this period in the Romanian East-Carpathians foreland. However, in front of the eastern Polish Carpathians, a normal fault system striking parallel to the orogen has been described by Krzywiec (2001). It was formed in post-Middle Badenian times and was active also during the Sarmatian (Fig. 4.10). Minor inversion in the hanging-wall of these normal faults occurred during the Late Badenian.

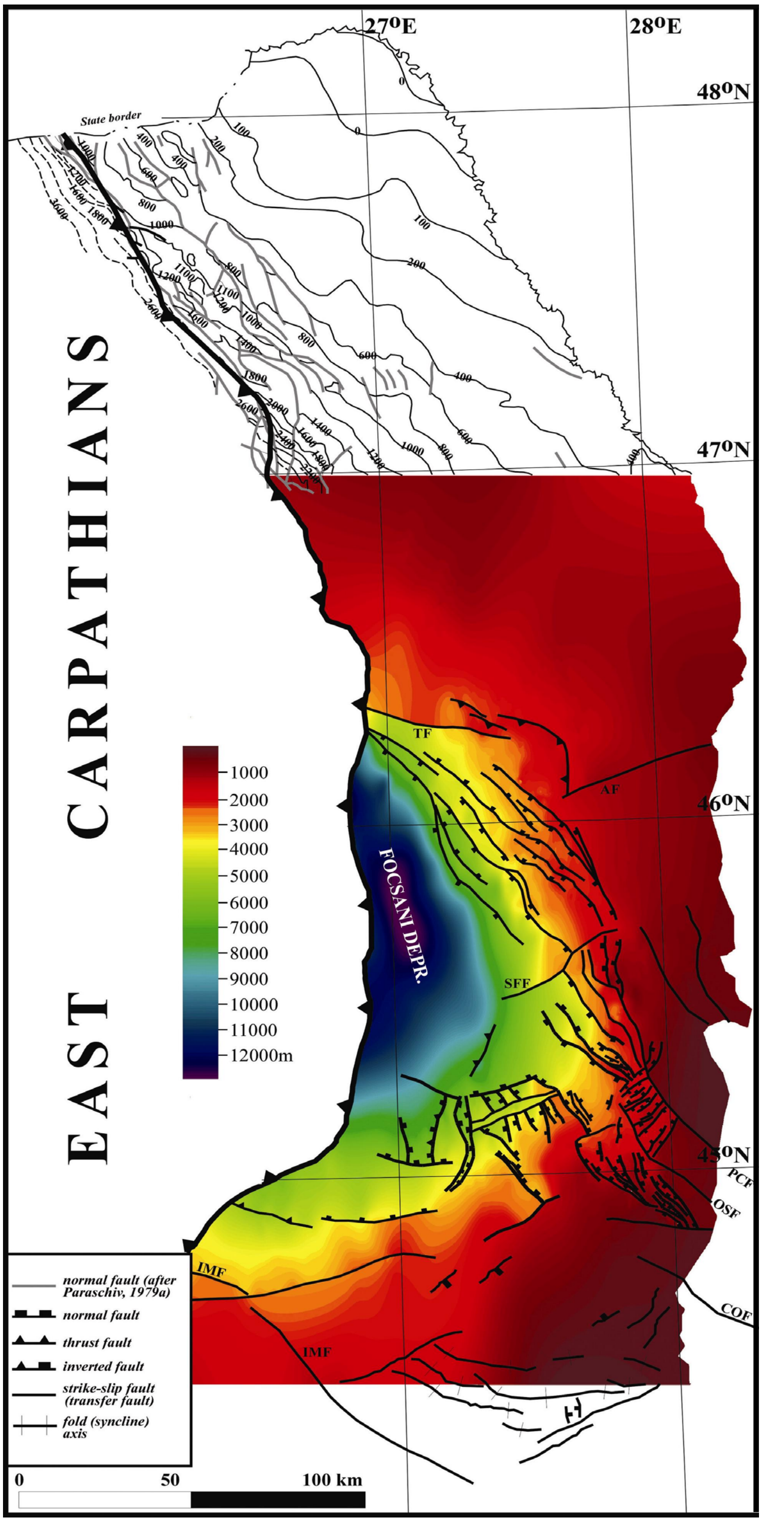
4.4.1.3. Basin tectonics

The overall Badenian subsidence pattern reflects a systematic increase in accommodation space and thus sediment thickness W-wards. Unfortunately, seismic data do not image very clearly Badenian sediments in the deeper parts of the Focşani Depression and it is, therefore, difficult to be more specific about how this W-wards thickening occurred.

Active structures are observed in the marginal parts of the basin and mainly consist of NW-SE trending normal faults found in the SE corner of the Focşani Depression and along its NE margin. The first group of faults defines several grabens (Figs. 4.7 and 4.8), which accommodate horizontal extension in the order of 10-20 km of a distance of ~100 km. How much of the observed subsidence can be ascribed to the associated NE-SW directed extension will be quantitatively tested in Chapter 5. Faults along the NE margin are apparent, but the vertical offset is in the order of hundreds of metres (Fig. 4.5) and there is no doubt that these faults cannot explain the bulk of the observed subsidence.

The cause of the Badenian extension is problematic because the Carpathians thrusting subsequently obscured the original configuration of the resulting basins, especially their continuation towards the W-NW of the Focşani Depression. A reconstruction of the structural setting as well as the full integration with the orogen tectonic history is required.

The northern foreland (i.e. East-European/Scythian platform) was gently tilted towards the orogen. Contractional structures from its southern margin as well as farther to the north (in front of the eastern Polish Carpathians; Krzywiec, 2001) can be documented suggesting that a mechanical coupling (sensu Ziegler et al., 2002) between the orogen and the strong East-European foreland exists.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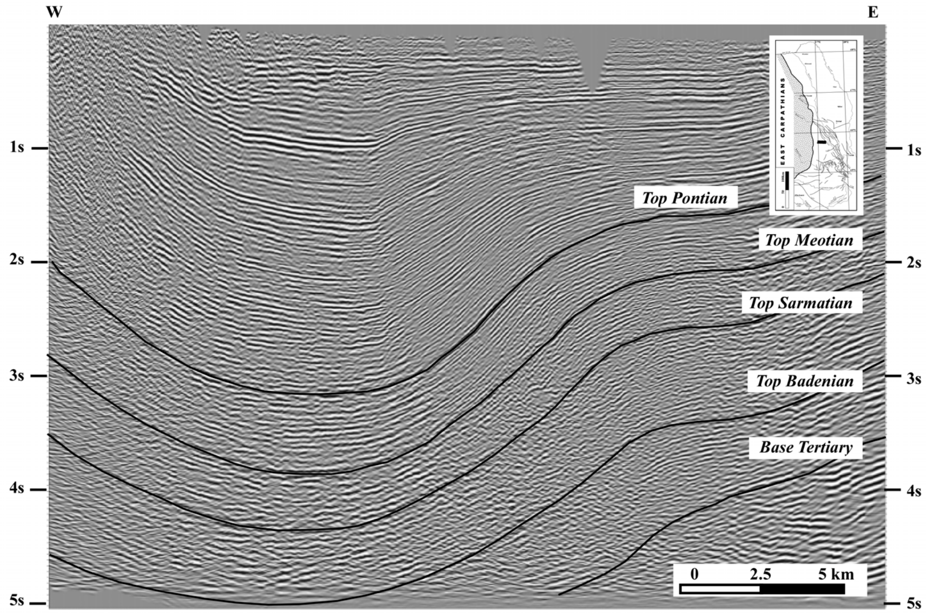
State border
EAST CARPATHIANS
FOCSANI DEPR.
TF
AF
SFF
PCF
OSF
COF
IMF
1000
2000
3000
4000
5000
6000
7000
8000
9000
10000
11000
12000m
normal fault (after Paraschiv, 1979a)
normal fault
thrust fault
inverted fault
strike-slip fault (transfer fault)
fold (syncline) axis
0
50
100 km

Figure 4.4 (previous page) *(**A**) Base Tertiary structural map. The reference datum is 100 m above sea level. The western margin of the map is represented by the reverse Caşin-Bisoca fault striking to the north up to approximately the Trotuş fault (e.g. Dicea, 1995), and farther north by the outcropping Pericarpathian fault (front of the Subcarpathian nappe). The Prut and Danube Rivers represent the eastern margin of the map. Colors have a depth scale. The northern part of the map (black contours and gray faults) is taken from Paraschiv (1979a). For the southernmost part of the map, only the structural pattern is shown. Abbreviations of the foreland faults are: AF Adjud fault; COF Capidava-Ovidiu fault; IMF Intramoesian fault; OSF Ostrov-Sinoe fault; PCF Peceneaga-Camena fault; SFF South-Focşani fault; TF Trotuş fault.*

Figure 4.4 (below) *(**B**) Seismic line across the Focşani Depression. Location in Figure 4.3. Note the semicircular shape of the basin and the impressive Neogene subsidence in the absence of significant fault systems. Vertical scale is two-way travel time.*



4.4.2. Late Miocene (Sarmatian)

4.4.2.1. Thickness and subsidence patterns

Sarmatian sediments are widespread over large areas, being found not only in the Focşani Depression but also farther to the north and south, on the East-European and Moesian platforms, respectively. On the whole, sediments are much less deformed and their thickness changes more gradually than the Badenian ones (e.g. Figs. 4.7 and 4.9). The Sarmatian deposits are represented by clastics and some calcareous intervals that indicate a shallow water environment, at least on the East-European/Scythin platform (Saulea, 1966, 1967; Saulea et al., 1969; Paraschiv, 1979a).

The area with the highest thicknesses of Sarmatian deposits is located in the western parts of the Focşani Depression where it forms a N-S trending basin, with thicknesses >5 km in its northern part (Fig. 4.11). Sarmatian subsidence rates are here ~1.3 km/Myr and decrease to 0.6 km/Myr in the central part and to 0.5 km/Myr farther south (Fig. 4.6). Towards the western boundary of the map shown in Figure 4.11, the base of the basin becomes steadily shallower. To the west of this boundary, Sarmatian sediments are partly incorporated in the Subcarpathian nappe or were mostly removed by the erosion that followed the Sarmatian (~11 Myr) shortening.

Sarmatian deposits up to 2.5 km thick are found outside the main depositional domain, on the East-European/Scythian platform, where they crop out (Saulea, 1966, 1967). The general trend is of increasing thicknesses towards the thrust belt (Fig. 4.11). Since the Sarmatian sequence is presently cropping out, the original thicknesses could be somewhat greater than the actual values. Over most of the East-European/Scythian platform the isopachs have a NW-SE trend, whereas in the southernmost part they become roughly N-S oriented.

To the east of Focşani Depression (along the North Dobrogea promontory) the Sarmatian sequence has thicknesses of a few hundreds of meters (Fig. 4.11). Here, the subsidence rates vary from 0.35 km/Myr to 0.13 km/Myr (Fig. 4.6, curve #4).

Sarmatian sediments are also found in front of the Carpathians Bend zone where they show subsidence rates of around 0.3 km/Myr (Fig. 4.6, curve #7), decreasing towards the SE. To the south of Focşani Depression, Sarmatian sediments are 1-1.2 km thick and progressively thin S-wards (Fig. 4.11). Following Badenian extension and subsequent mild inversion, the first Sarmatian reflectors onlap the erosional unconformity that marks the end of the Badenian deposition (Fig. 4.7). Also, the Sarmatian sequence extends farther south than the Badenian pinch-out limit (Figs. 4.12A and B).

Thick prograding sedimentary bodies form the Middle-Late Sarmatian basin fill, with transport direction from north to south, i.e. along the foredeep axis (Fig. 4.13; Negulescu, 2001). These prograding bodies are not confined only to areas near the Carpathians thrust front, but extend also over a much larger area to the east. A sequence with oblique-tangential clinoforms and mound-shaped sedimentary bodies to the distal part is clearly identified between ~0.5-1 s in Figure 4.13. From the elevation difference between the topsets and bottomsets, it appears that the progradation took place in a basin as deep as ~400-500 m (~400 ms). The age of this sequence is Middle Sarmatian (Negulescu, 2001), that is, the deepening of the basin can be correlated with the emplacement of the Subcarpathian nappe over the foreland (Săndulescu, 1984, 1988).

4.4.2.2. Active structures

The Sarmatian was a time of limited deformation in the Focşani Depression and only few active structures have been identified (Fig. 4.11). In the northern foreland (East-European/Scythian platform) large-scale tilting and major subsidence took place (Figs. 4.14A and B). Normal faulting stretching parallel to the orogen, mainly W-SW-dipping, accompanied the tilting of the East-European/Scythian platform. The magnitude of faulting is minor in the southernmost part (e.g. Fig. 4.14A) and increases progressively to the NW along the thrust belt (Fig. 4.14B; see also Răileanu et al., 1994). To the north of the Focşani Depression, the major active structure was the Trotuş fault, which accommodated the subsidence of the southern block (Fig. 4.11). The Trotuş fault is imaged as a negative flower structure (Fig. 4.15A) with a sinistral sense of movement determined from kinematic studies carried out in the orogen along its prolongation (cf. Maţenco, 1997; Maţenco and Bertotti, 2000).

To the south, a cluster of NE-SW striking normal faults separated by transfer zones is found mainly between the Peceneaga-Camena and Ostrov-Sinoe faults (Fig. 4.11). Peceneaga-Camena fault is interpreted as a wide shear zone between Moesia and North Dobrogea orogen (Fig. 4.15B). Narrow NE-SW-oriented basins are interpreted as pull-aparts related to a dextral movement along the two main faults. However, taking into account the dimension of the pull-apart basins, the displacement along the master strike-slip faults should have been minor, probably up to a few kilometres. Other faults with associated flower structures are observed to the east of the Peceneaga-Camena shear zone (Fig. 4.15B). Since they parallel the Peceneaga-Camena fault, a dextral sense of movement is proposed. Neither the Peceneaga-Camena nor the other faults can be followed along strike to the north of the S-Focşani fault (Fig. 4.11). The S-Focşani fault thus represents the boundary between a W-wards tilting region to the north (basically undeformed) and a region experiencing normal and strike-slip faulting to the south. To the south of the interpreted pull-apart basins, some of the Badenian normal faults were apparently reactivated (Figs. 4.8 and 4.11), and might correlate with the structures found presently in Dobrogea at the surface (Hippolyte, 2002).

In the southwestern part of the studied area, dextral movements took place along the Intramoesian fault (Figs. 4.11 and 4.15C). This sense of movement is supported by the NW-SE trending folds from the eastern block as well as by other small-scale associated structures (Tărăpoancă, 1996). Transpression is recorded along its WNW-ESE trending segment resulting in a SE-trending reverse fault near the Carpathians Bend, its western edge being covered by the Carpathians structures (Fig. 4.11).

Between the Intramoesian and Peceneaga-Camena faults, ENE-WSW oriented faults were active possibly as strike-slip faults with a sinistral sense of movement (Figs. 4.11, 4.12A and B). These faults were probably responsible for the minor erosion occurring at the end of the Sarmatian as observed in Figure 4.12A. This unconformity is developed only locally, in the easternmost former Badenian basin.

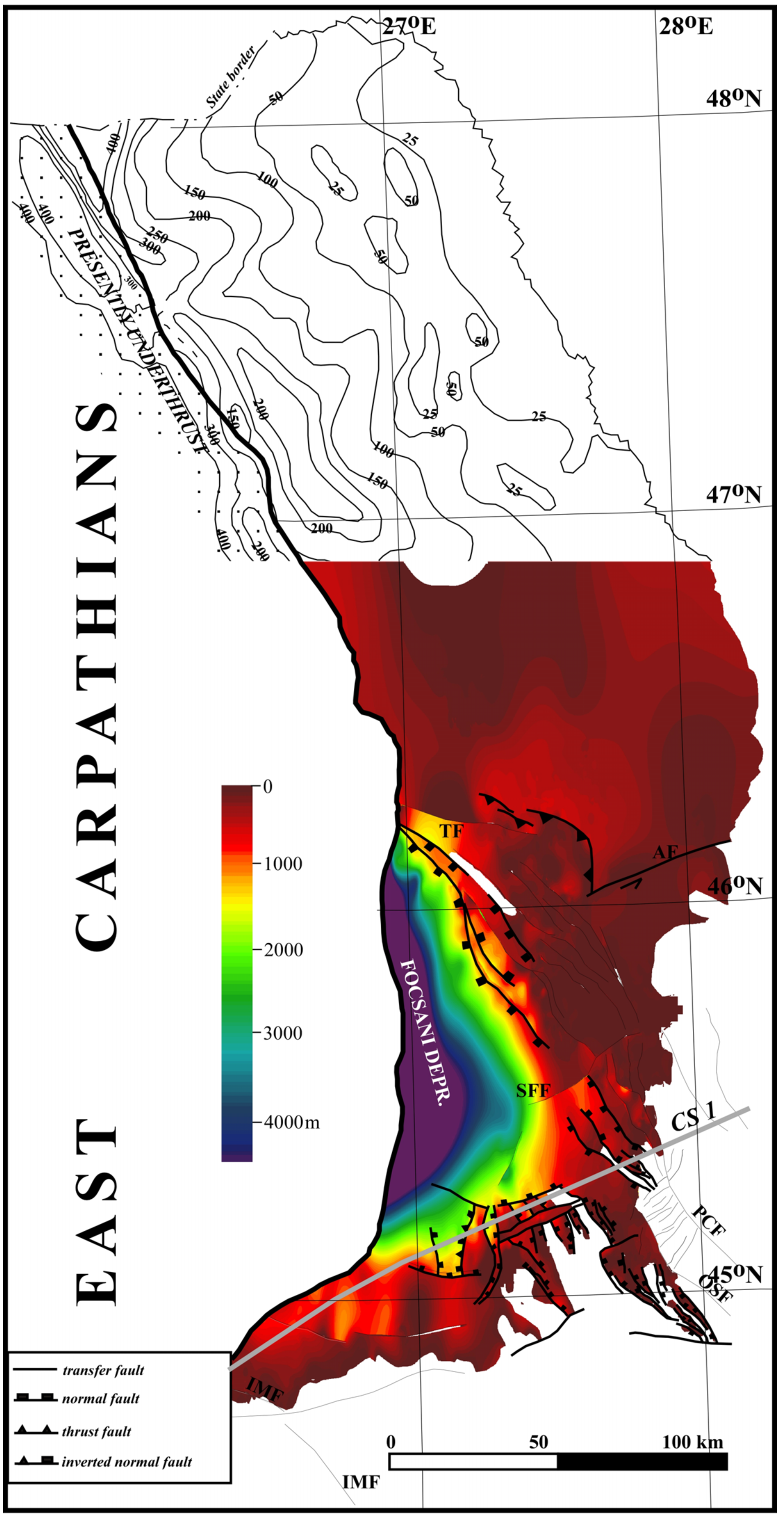
4.4.2.3. Basin tectonics

At the Badenian/Sarmatian transition, major changes in the evolution of the Focşani Depression occurred. The geographic distribution of sediment thicknesses changed. During the Sarmatian, the main depocenter became narrower and more elongated in N-S direction (Fig. 4.11). Subsidence rates in this central domain are slightly decreased with respect to the Badenian ones. This area is also characterized by a thicknesses decrease towards the west, pointing to a synclinal geometry of the Sarmatian basin.

At the same time, the basin expanded to the east and sediments were deposited on previously stable areas up to several hundred of kilometres to the east of the Carpathian front. This is the case for the East-European/Scythian platform (e.g. Figs. 4.14A and B) and, farther to the south, the region of the Intramoesian fault (e.g. Fig. 4.12B). In these areas, thickness changes are very gradual, indicating a regional control on subsidence. Subsidence curves outside the depocentral areas clearly follow this trend showing a significant increase in the rate of accommodation space creation (Fig. 4.6).

An important difference between the East-European/Scythian platform and Moesia must be mentioned: there is no Badenian/Sarmatian interruption in sedimentation in the northern foreland, moreover, an increase in subsidence rates is observed. Here, the Badenian/Sarmatian boundary is correlated with a facies change, from predominantly evaporitic in the Badenian to dominantly clastic in the Sarmatian (e.g. Paraschiv, 1979a; Ionesi, 1994; Negulescu, 2001).

Figure 4.5 (next page) *Isopach map of the Badenian sequence. No decompaction correction was applied. The northern map (black contours) is taken from Paraschiv (1989). Highlighted structures are those active during the Badenian. Gray line (CS 1) corresponds to the cross-section used in extensional modeling in Chapter 6. Abbreviations as in Figure 4.4.*



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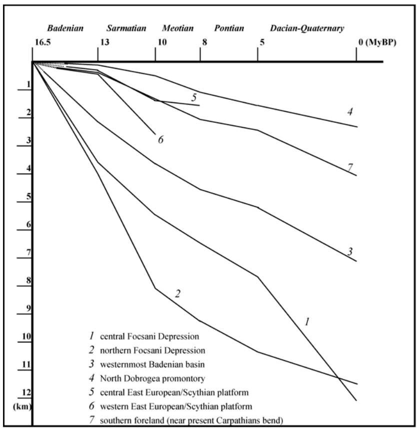


Figure 4.6 *Total subsidence curves. Location in Figure 4.3. Paleo-bathymetry is taken as zero meters considering that the sedimentation generally occurred in shallow water environments (Fig. 2.7; Saulea et al., 1969). Also, no correction for sedimentary load was applied.*

By contrast, an erosional unconformity represents the Badenian/Sarmatian boundary in Moesia, south of the S-Focşani fault (e.g. Fig. 4.7).

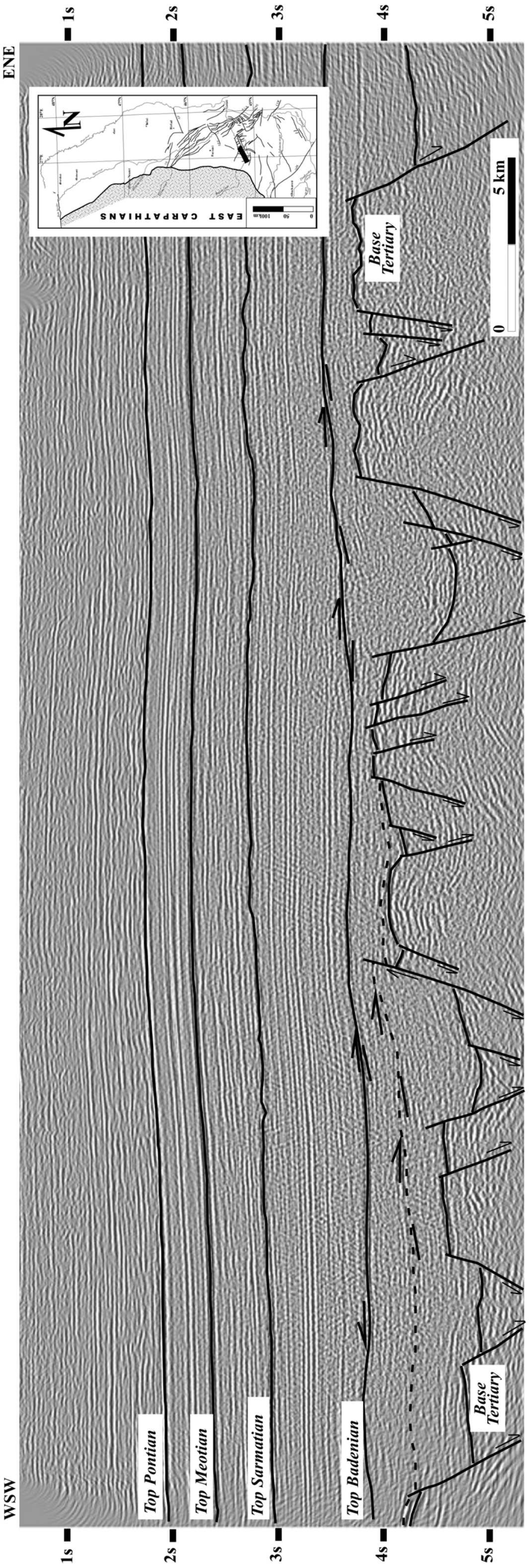
Orogen-ward tilting of Moesia (except for the area of the Focşani Depression) was less intense than of the East-European/Scythian platform, as reflected by the sediment thicknesses (compare for instance, Fig. 4.12B with Fig. 4.14A, respectively). The emplacement of the Subcarpathian nappe on the foreland controlled this tilting. After this contractional event (Middle Sarmatian, ~11 Myr), no significant deformations affected the central-northern Carpathians, i.e. to the north of Trotuş fault.

Flexure-related normal faulting along the internal margin of the East-European/Scythian platform is replaced to the south of the Focşani Depression by mainly strike-slip faulting. With apparently only minor displacements in the easternmost part of Moesia, the magnitude of dextral motion along NW-SE trending faults seems to increase to the west (to the Intramoesian fault).

The tectonic activity documented on the Trotuş and Intramoesian faults in the north and south, respectively, confirms that the orogenic wedge confined between these two major alignments experienced ESE-ward escape during the Late Sarmatian as proposed by Maţenco and Bertotti (2000).

Figure 4.7 (next page, up) *Interpreted seismic line from the southern foreland showing the Badenian extensional basins. Location in inset (Fig. 4.3). Note that a small inversion occurred to the end of Badenian in the western part (dashed line marks the onset of inversion with onlapping reflectors above). Also, the Badenian/Sarmatian boundary represents an erosional unconformity (thicker black arrows denote onlap terminations and black bars denote erosional truncations). Vertical scale is two-way travel time.*

Figure 4.8 (next page, down) *Interpreted seismic line across the Badenian basins from the southern foreland near the SE edge of the extended region. Location in inset (Fig. 4.3). Here, the basins are narrower and shallower than those formed to the north (compare with Figure 4.7). By contrast, the syn-rift phase continued here until the Sarmatian and even into the Meotian. A general N-wards tilting occurred after the Pontian (note the wedge shape of the post-Pontian sedimentary sequence). Vertical scale is two-way travel time.*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WSW
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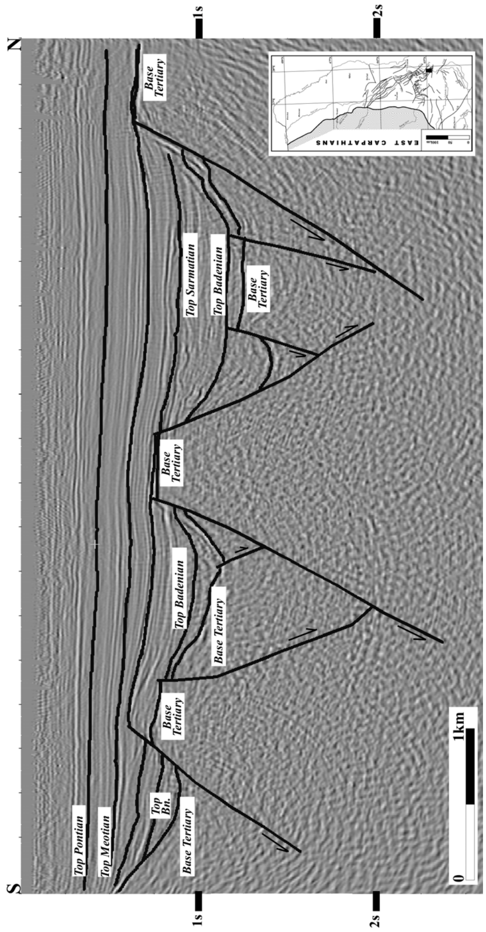
2s
3s
4s
5s
Top Pontian
Top Meotian
Top Sarmatian
Top Badenian
Base Tertiary
0
5 km
EAST CARPATHIANS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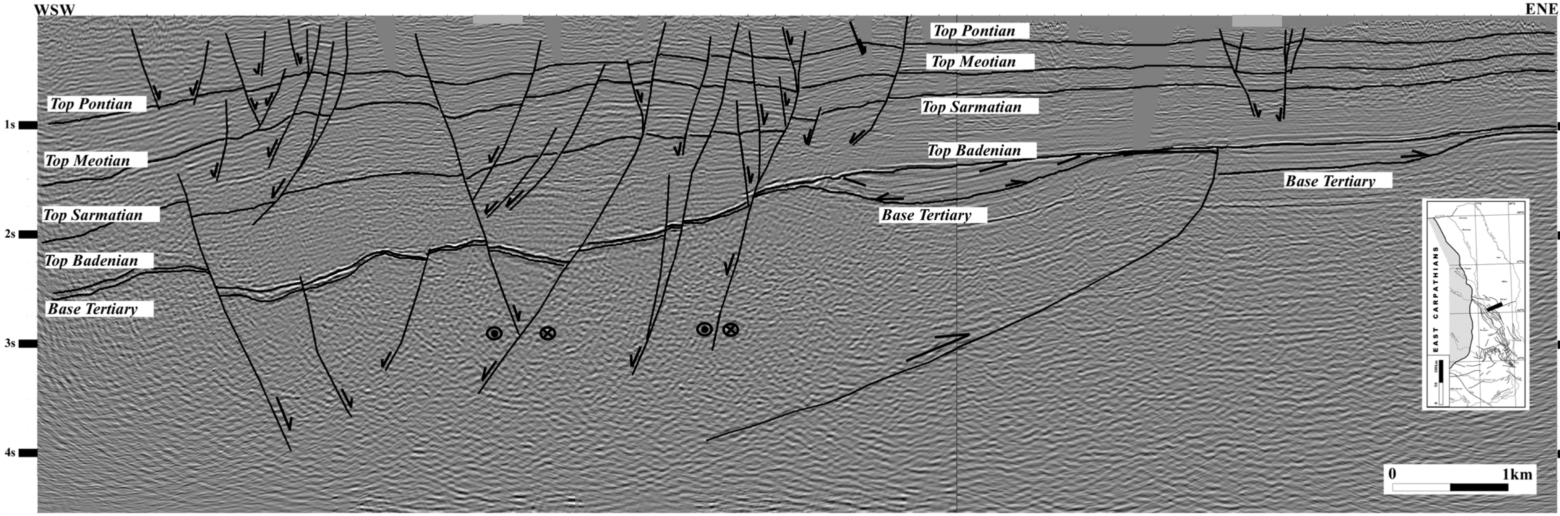


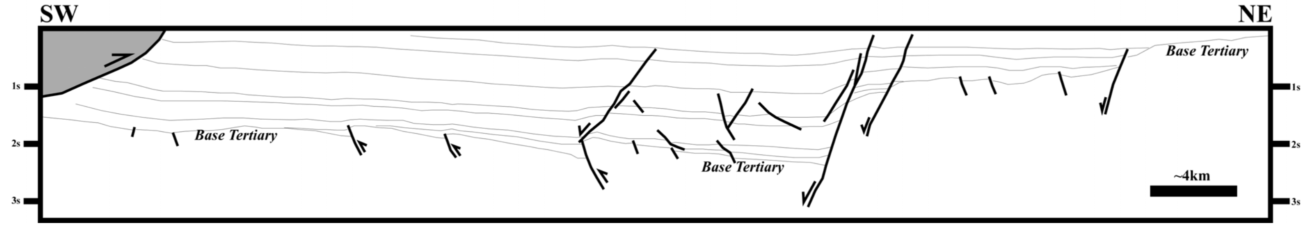


Figure 4.9 (up) *Interpreted seismic line from the northern foreland. Location in inset (Fig. 4.3). A thrust fault is interpreted to the eastern part of the line. Piggyback and flexural-associated basins developed during the Badenian. The other structures were mainly formed during the Pliocene (after Pontian). These features are interpreted as normal faults but some negative flower structures could be observed as well.*

Figure 4.10 (down) *Geological interpretation of a seismic line from the Polish Carpathians close to the juction between West- and East-Carpathians (after Krzywiec, 2001). The sedimentary section overlying the pre-Tertiary base foredeep comprises Lower Badenian-to-Lower Sarmatian deposits. Both normal and reverse faults are evident. The onset of faulting is post-Middle Badenian. The major normal fault resulted from flexural extension and reactivation of inherited structures, whereas the reverse faults are produced due to compression related to the Carpathians collision, according to Krzywiec (2001). For location, see Krzyywiec (2001). In both lines, vertical scale is two-way travel time.*

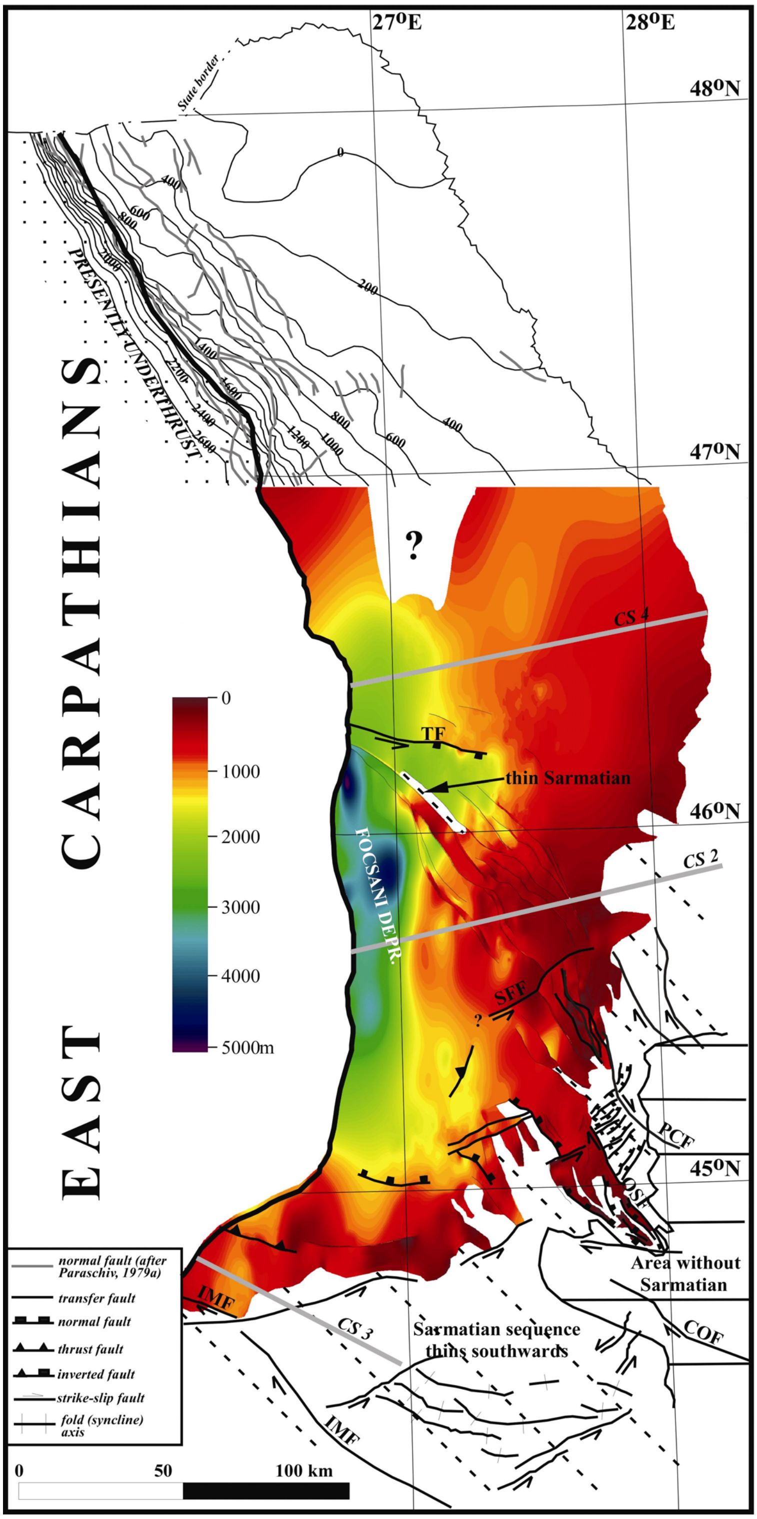

27°E
28°E
48°N
47°N
46°N
45°N
State border
PRESENTLY UNDERTHRUST
EAST CARPATHIANS
CS 4
CS 2
CS 3
TF
thin Sarmatian
FOCSANI DEPR.
SFF
PCF
COF
IMF
Area without Sarmatian
Sarmatian sequence thins southwards
0
1000
2000
3000
4000
5000m
normal fault (after Paraschiv, 1979a)
transfer fault
normal fault
thrust fault
inverted fault
strike-slip fault
fold (syncline) axis
0
50
100 km

Figure 4.11 (previous page) *Isopach map of the Sarmatian sequence. No decompaction correction was applied. The northern map (black contours and gray faults) represents the difference between the maps from the Figures 4.4 and 4.5 (Paraschiv, 1979a, 1989, respectively). Highlighted structures are those that were active during the Sarmatian. Gray lines (CS 2, 3 and 4) represent the cross-sections used as reference in the flexural modeling in Chapter 6. Abbreviations as in Figure 4.4.*

4.4.3. Latest Miocene (Meotian)

4.4.3.1. Thickness and subsidence patterns

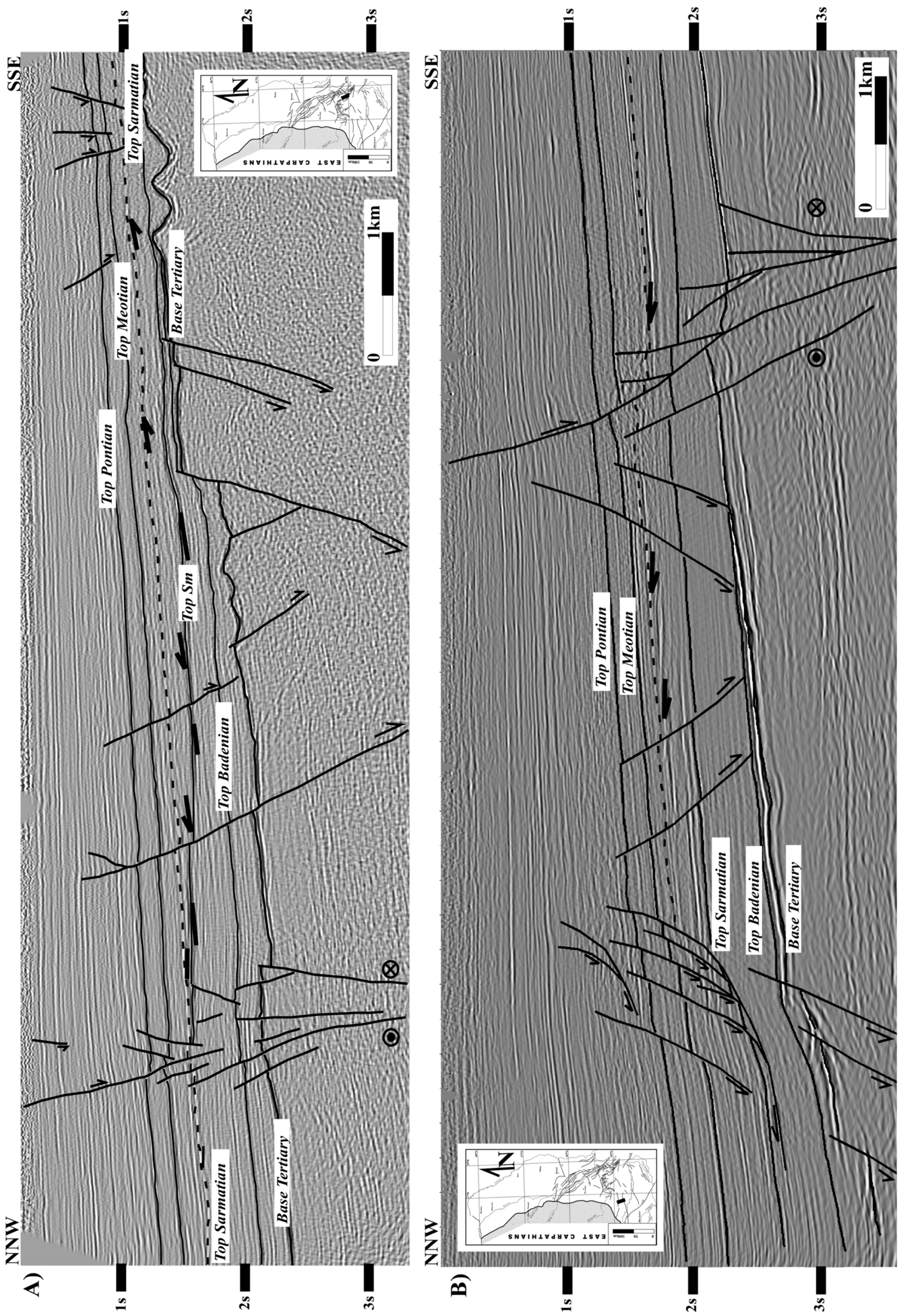
Meotian deposits (Fig. 4.16A) are widespread over the entire area of the Focşani Depression. Depocentres are identified south of the Trotuş fault (roughly coinciding with the Sarmatian ones) and, farther to the south, in front of the Carpathian Bend zone. Thicknesses reached in these sectors are 1.5-1.6 km. Isopachs outlining these depocentres trend roughly N-S. Meotian subsidence rates in the Focşani Depression are around 0.6 km/Myr in the northern part and 0.5 km/Myr in the central part (Fig. 4.6, curves #2 and 1, respectively).

On the East-European/Scythian platform, the area of sedimentation was drastically reduced compared to the Sarmatian one (Fig. 4.16A). Here, Meotian deposits are thin (e.g. Fig. 4.14) and cover only a limited region to the north of the Trotuş fault. They are typically 0.7-0.9 km thick and decrease to zero SE-wards. Subsidence rates are correspondingly low and are at best 0.1 km/Myr (Fig. 4.6, curves #5 and 6). Since some Meotian deposits were subsequently removed by erosion, the actual subsidence rate may have been slightly higher.

Outside the main deposition centres, Meotian sediments are found over most of Moesia and the North Dobrogea orogen (promontory) with very gently changing thicknesses. From the Focşani Depression eastwards, the Meotian sedimentation rate is around 0.27 km/Myr (Fig. 4.6, curve #4). Similar values of a few hundred metres per million years were derived for the southern regions.

The Sarmatian southwards sediment progradation continued during the Lower Meotian over almost the entire foreland to the south of Focşani Depression, the end of this sedimentation pattern being marked by an extensive toplap surface (Fig. 4.12A). The SW border of this basin is formed by the Intramoesian fault. Only in the easternmost part of Moesia, did Lower Meotian sediments prograde from SE to NW (Fig. 4.12A), with Dobrogea probably as a source area. The end of this progradation is represented by a toplap surface and locally by erosional truncation.

Figure 4.12 (next page) *Interpreted seismic lines from the southern foreland. Location in inset (Fig. 4.3). A strike-slip fault is interpreted in each line, to the north in* ***A*** *and to the south in* ***B*** *(note the occurrence of both normal and inverse offsets and the change in layer thicknesses across the some faults). Post-Pontian normal faults (planar and listric) are also interpreted. The dashed black line within the Meotian sedimentary sequence in both lines represents a synchronous toplap surface and the thicker black arrows correspond to toplap stratal terminations. Note that in* ***A****) the progradation is from S to N whereas in* ***B****) the sense is reversed. Erosional truncations (black bars) are observed in* ***A****) at Top Sarmatian and Top Meotian deltaic sequence and could be related to the tectonic activity of the strike-slip fault. No such unconformities are seen in* ***B****). Both lines show important N-wards tilting, which started in the Pontian and increased during the Pliocene (note the thickening of these sedimentary sequences to the north). Vertical scale is two-way travel time.*


A)
NNW
SSE
1s
2s
3s
Top Sarmatian
Top Meotian
Base Tertiary
Top Pontian
Top Sm
Top Badenian
Top Sarmatian
Base Tertiary
0
1km
EAST CARPATHIANS
N
B)
NNW
SSE
1s
2s
3s
Top Pontian
Top Meotian
Top Sarmatian
Top Badenian
Base Tertiary
0
1km
EAST CARPATHIANS
N

4.4.3.2. Active structures

Little localized deformation occurred during the Meotian, mainly consisting of the reactivation of the synthetic and antithetic Badenian normal faults along the NE margin of the Focşani Depression (Figs. 4.9 and 4.16A). Farther to the east, new NW-trending normal faults formed. Vertical offsets increase towards the SE reaching a few hundred meters.

During the same time span, the S-Focşani fault acted as a transfer fault. The Peceneaga-Camena and some of the associated faults were reactivated either as normal or strike-slip (e.g. Figs. 4.8 and 4.15B respectively). In the SE part of the Focşani Depression, a localized erosional unconformity formed at the end of the Early Meotian and is probably related to the reactivation of one such strike-slip fault (Fig. 4.12A).

In the SW part of the studied area, the Intramoesian fault experienced dextral strike-slip movements (Figs. 4.15C and 4.16A; also Tărăpoancă, 1996) with a component of normal displacement increasing towards the north. On the whole, all active structures trend NW-SE and affect the eastern margin of the Focşani Depression.

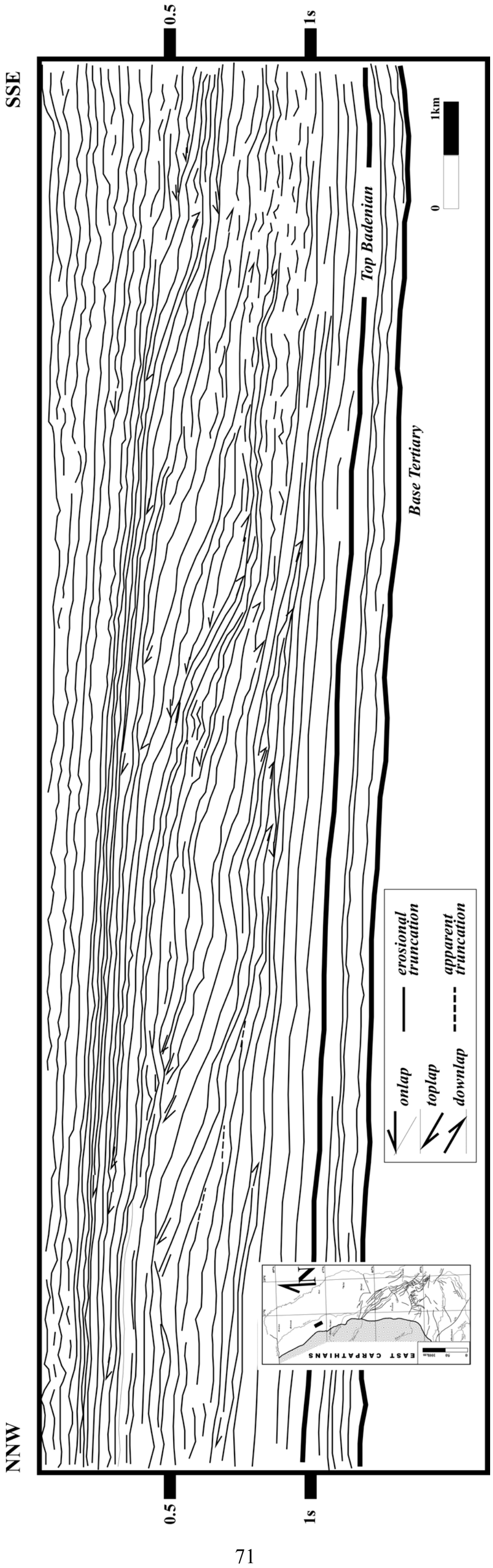
4.4.3.3. Basin tectonics

The Meotian subsidence pattern partly resembles the one reconstructed for the Sarmatian. The main depocenter area still forms an N-S elongated stripe in front of the already structured Carpathian belt and is flanked to the east by a wide domain with persistent sedimentation and fairly regular thicknesses (Fig. 4.16A).

Continuing the existing trend, subsidence rates in the depocentral areas are lower than in the Sarmatian (curves #1 and 2 in Fig. 4.6). In contrast, a slight increase in subsidence rates is observed in the eastern regions (curve #4 in Fig. 4.6).

The most obvious changes with respect to the Sarmatian subsidence pattern are the cessation of subsidence north of Trotuş fault (e.g. Fig. 4.14 and curve #5 in Fig. 4.6) and the development of an area with reduced subsidence in the westernmost sectors of the Focşani Depression (Fig. 4.16A). As in the Sarmatian, Meotian sediments are basically undeformed and the generation of accommodation space reflects regional downwarping of the basin floor, although some normal faulting is documented along the eastern margin of the Focşani Depression (e.g. Fig. 4.9).

Figure 4.13 (next page) *Interpreted seismic line from the northern foreland, roughly parallel to the East Carpathians (modified after Negulescu, 2001). Location in inset (Fig. 4.3). Note the large-scale Sarmatian deltaic progradation along the foredeep, from NNW to SSE. Sediments prograded continuously to the SSE indicating that the accumulation space was always in the Focşani Depression. Some forced regressions and transgressive system tracts can be also observed. Compare this sedimentation setting with the thin, rather uniform, parallel-bedded Badenian sequence. Vertical scale is two-way travel time.*


SSE
NNW
0.5
1s
0.5
1s
Top Badenian
Base Tertiary
0
1km
onlap
toplap
downlap
erosional truncation
apparent truncation
EAST CARPATHIANS

4.4.4. Latest Miocene (Pontian)

4.4.4.1. Thickness and subsidence patterns

During the Pontian, the older depocenters immediately south of the Trotuş fault and in front of the Carpathian Bend Zone were abandoned and the maximum accommodation space was created in the central part of the investigated area. In the central parts of the Focşani Depression (Fig. 4.16B) thicknesses reach 1.5-1.6 km and isopachs are mainly NNW-SSE oriented. Sediment thicknesses decrease towards the west. The Pontian subsidence rate in the Focşani Depression is around 0.4-0.45 km/Myr in the central domains (curve #1 in Fig. 4.6) decreasing towards the north (curve# 2 in Fig 4.6).

Pontian sediments are missing to the north of the Trotuş fault. Thicknesses regularly decrease from the main depocentral area towards the east (North Dobrogea promontory) and the sediments eventually pinch out. The depositional limit of the Pontian oversteps E-wards the Meotian basin margin (e.g. Fig. 4.15B).

Subsidence rates in the area are in the order of 0.1-0.2 km/Myr (curve #4, Fig. 4.6). In the southern foreland (south of Focşani Depression), thicknesses decrease rather gradually from the depocentral areas towards the south (Figs. 4.8, 4.12A and B) and very few Pontian sediments are observed in the region of the Intramoesian fault. Subsidence rates are again in the order of 0.1-0.2 km/Myr (Fig. 4.6, curves #3 and 7). During the Sarmatian and Meotian, the central-northern East-Carpathians (main source area) was probably characterized by a high

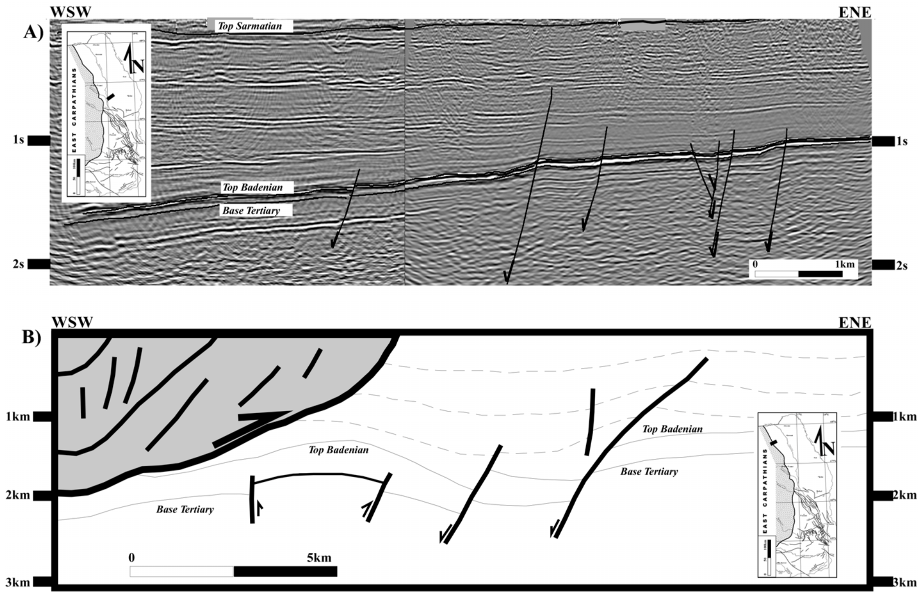


Figure 4.14 *Interpreted seismic lines from the northern foreland, roughly perpendicular to the East Carpathians. Location in inset (Fig. 4.3). Line* ***B****) is taken from Negulescu (2001). Vertical scale is two-way travel time in* ***A****) and depth (kilometres) in* ***B****). In both lines, the Sarmatian sequence is wedge-shaped, indicative of a typical foredeep basin infill. The Badenian sequence is thin and thickens very gently to the west. Note the increase in normal faults offset from* ***A*** *(less than 100 m) to* ***B****. The inverse faults from* ***B****) interpreted by Negulescu (2001) might be compared with those from Figure 4.10 (West-Carpathians foreland).*

relief and supplied detritus to large S-wards prograding deltas (Figs. 4.12B and 4.13). However, during the Pontian, the elevation of this source area was probably much lower as the lithology of Pontian deposits is dominantly pelitic (e.g. Ionesi, 1994; Jipa, 1997).

4.4.4.2. Active structures

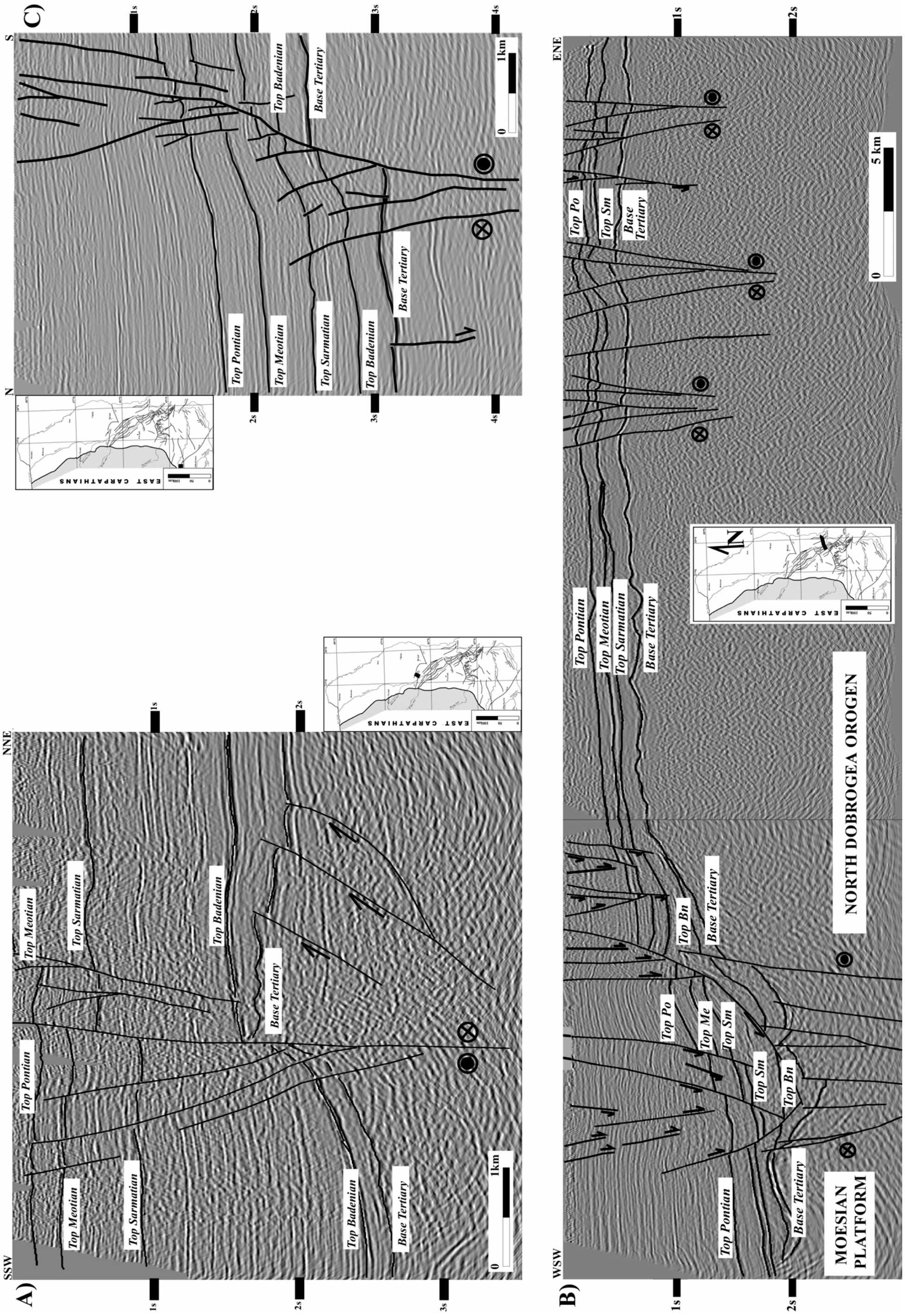
The Pontian is the period with the least tectonic activity, and despite ongoing vertical movements, no important structures formed. Only few normal faults continued to be active along the E-SE margin of the Focşani Depression (Fig. 4.16B).

4.4.4.3. Basin tectonics

The Pontian was a time of significant changes in the subsidence pattern of the Focşani Depression and marks the transition to the presently active pattern of vertical movements. The main depocentre moved a few tens of kilometres to the SE with respect to its position during Meotian times (compare Figs. 4.16B with A). Consequently, (1) the neighborhood of the Trotuş fault ceased to subside and, (2) the axis of the Focşani Depression moves away from the Carpathians. This means that maximum foredeep subsidence occurs 10-to-25 km to the front of the tip of the orogenic load and that the basin shallows towards the thrust belt via a stable to uplifting area.

Subsidence rates in the subsiding areas continue to decrease with respect to previous periods. The elevated NNW-SSE trending area in the western portion of the Focşani Depression persisted where Pontian sediments are very thin to missing. This zone parallels the main depocentre but is slightly oblique with respect to the Carpathians trend. No sedimentation occurred on the East-European/Scythian platform, which behaved as a stable or uplifting region. By contrast, Moesia was regionally tilted towards the orogen, thus creating available accommodation space in the absence of active structures.

Figure 4.15 (next page) *Platform-bounding faults: Trotuş (**A**), Peceneaga-Camena (**B**) and Intramoesian (**C**). Locations in inset (Fig. 4.3). Note the important offset of the foredeep base (pre-Tertiary) across all these faults.*
A) *The Trotuş fault separates the Scythian platform to the north from Moesia to the south (cf. Săndulescu and Visarion, 1988) and is interpreted as a sinistral strike-slip fault (sense of displacement inferred by Maţenco, 1997 and Maţenco and Bertotti, 2000). Important vertical offset was produced during the Sarmatian. Young deformations (post-Pontian) can be noticed as well. In the uplifted block, Badenian inverse faults belong to the same thrust system (splays of the main fault) identified in Figure 4.9.*
B) *The Peceneaga-Camena fault separates Moesia to the west from the North Dobrogea orogen to the east and is interpreted as a dextral shear zone (sense of movement discussed in text). Downward movement of Moesia relative to the North Dobrogea orogen was produced especially in the Pliocene-Quaternary, as proven by the W-wards thickening of the post-Pontian sequence. Inside the North Dobrogea orogen, negative and positive flower structures are interpreted to be present.*
C) *The Intramoesian fault separates Moesia into two blocks with different crystalline basements and other characteristics, such as heat flow, crustal and lithospheric thicknesses (Chapter 2). It is interpreted as a dextral fault active during most of the Neogene (see the text).*


A)
SSW
NNE
Top Pontian
Top Meotian
Top Sarmatian
Top Badenian
Base Tertiary
1km
EAST CARPATHIANS
B)
WSW
ENE
MOESIAN PLATFORM
NORTH DOBROGEA OROGEN
Top Po
Top Me
Top Sm
Top Bn
Base Tertiary
5 km
C)
N
S
1s
2s
3s
4s

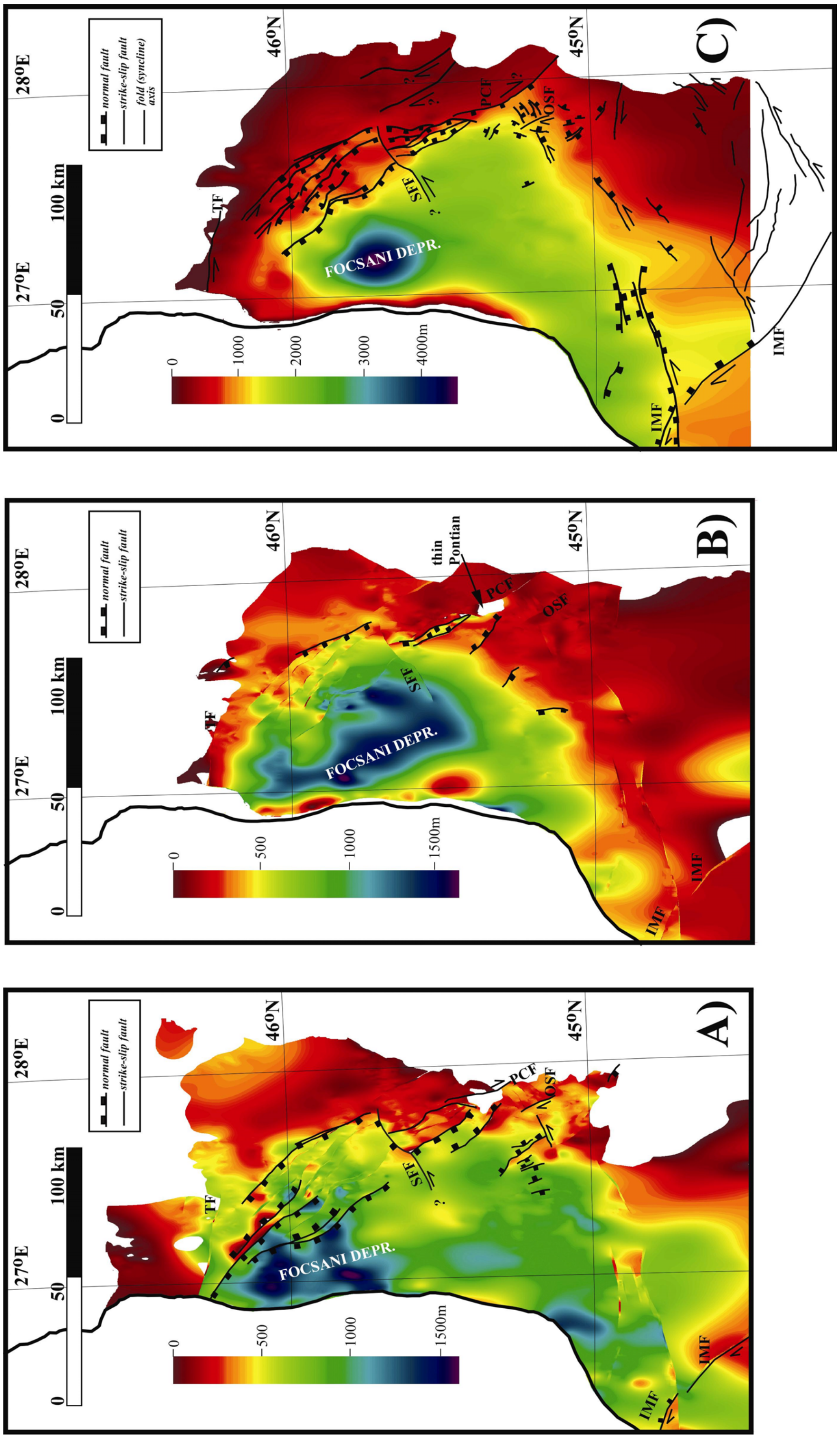

C)
B)
A)
FOCSANI DEPR.
normal fault
strike-slip fault
fold (syncline) axis
0
50
100 km
27⁰E
28⁰E
46⁰N
45⁰N
TF
SFF
PCF
OSF
IMF
thin Pontian
0
1000
2000
3000
4000m
0
500
1000
1500m

Figure 4.16 (previous page) *Isopach maps of the Meotian (**A**), Pontian (**B**) and Pliocene-Quaternary (**C**) sequences. No decompaction correction was applied. In **C**), the present relief has not been taken into account; instead, a reference datum of 100 m above sea level is used. Note that the scale of color-coding is different in the 3 panels. Highlighted structures are those active at the specified time interval. Abbreviations as in Figure 4.4.*

4.4.5. Pliocene-Quaternary

4.4.5.1. Thickness and subsidence patterns

During Pliocene-Quaternary times, the central sectors of the Focşani Depression subsided strongly. Pliocene-Quaternary deposits reach a maximum thickness of almost 4.5 km in a subcircular area in the centre of the Focşani Depression (Fig. 4.16C). Outside this area, isopachs between 2,500 and 1,000 meters have a distinct, linear NNW-SSE trend in the east and NE-SW in the south. High subsidence rates up to 0.86 km/Myr were obtained for the central parts of the Focşani Depression (curve #1 in Fig. 4.6). The depocentre is flanked westwards by an area of reduced or no sedimentation (Fig. 4.16C).

Moving away from the main depocentre, Pliocene-Quaternary sediments thin towards the north and reach zero near the Trotuş fault. Subsidence rates are equal to few hundred metres per million years (e.g. curve #2 in Fig. 4.6). Towards the east, coeval sediments are spread over the entire area up to the Peceneaga-Camena fault. They are 1 km thick on the North Dobrogea promontory and around 2 km in the domain between the Intramoesian and Peceneaga-Camena faults (Figs. 4.15B and 4.16C). Subsidence rates on the North Dobrogea promontory are around 0.15 km/Ma (curve #4 in Fig. 4.6).

The thickness pattern reveals the roughly uniform NW-wards tilting of the southern foreland (Figs. 4.8, 4.12A, B and 4.16C). To the SSE of the Focşani Depression, the thicknesses gently decrease. An abrupt decrease in thickness in order of several hundred of metres is observed across the Intramoesian fault (Figs. 4.15C and 4.16C).

In terms of lithofacies, the Miocene (Pontian)/Pliocene boundary represents the change from pelitic to (progressively) coarser sediments (e.g. Paraschiv, 1979a; Ionesi, 1994).

4.4.5.2. Active structures

Widespread faulting occurred during Pliocene to Quaternary times, compatible with the large vertical movements recorded in the sedimentary succession. The NW-SE trending system of normal faults at the NE margin of the Focşani Depression was reactivated (Figs. 4.9 and 4.16C) in some cases with negative flower structures associated with presumably minor sinistral strike-slip movements. Vertical offsets decrease from NW to SE. Sinistral movements occurred along the Trotuş (Fig. 4.15A) and S-Focşani faults (Fig. 4.16C).

In the southern domains, the basin experienced further NNW-ward tilting, partly related to movements along the Intramoesian and Peceneaga-Camena faults (Figs. 4.15B, C and 4.16C). It seems that apart from the observed downwards displacement, a component of strike-slip occurred as well, especially on the Intramoesian fault (note the flower structure affecting the post-Pontian sequence in Figure 4.16C). A dextral movement is inferred, based on the pattern of contemporaneously active NW-trending folds and associated faults. ENE-WSW trending faults, such as those imaged in Figures 4.12A and B, were reactivated. The NW trending folds as well as the newly activated and relatively small NE-trending normal fault system seem to be related to sinistral displacements along the southern ENE-trending strike-slip faults (Fig. 4.16C). Taking into account the magnitude of associated folding/faulting, the amount of displacement along these sinistral faults must be small, probably in order of a few kilometres at best.

A dextral reactivation of the Peceneaga-Camena and the other parallel faults of the North Dobrogea orogen, shown in Figure 4.15B, is proposed. At least, the cross-sectional shape of the fault system from the North Dobrogea orogen (typical flower structures, both positive and negative, with changes in layers thickness as well as normal and reverse offsets on the same fault) strongly indicates strike-slip deformation. However, the sense of movement is not clear and it is tentatively interpreted as a dextral displacement. The boundary between the northern, sinistral domain and the southern, dextral one is represented by the S-Focşani fault (Fig. 4.16C). On the whole, this fault may have played a similar role to the Trotuş fault in separating a strongly subsiding region to the north (Focşani Depression) from a more gently inclined region to the south (e.g. Fig. 4.4). Unfortunately, W-wards prolongation of the S-Focşani fault cannot be documented.

4.4.5.3. Basin tectonics

The northern part of the Focşani Depression, which had strongly subsided during the Pontian, continued to do so during Pliocene-Quaternary times. More than 4 km of sediments were deposited in this area, in response to large wavelength vertical movements (e.g. Figs. 4.12A and B). Sedimentation rates increase in the central part of Focşani Depression with respect to previous time frames (Fig. 4.6). On the NE flank of the depression isopachs display a distinct NW-SE trend, parallel to the major faults in this area. Along the western margin of the Focşani Depression, the uplifting area, which came into evidence during Pontian times, developed further and was associated with steepening and tilting of pre-Pliocene beds. At the southern part of the investigated zone, subsidence affected previously stable areas, involving large domains to the NE of the Intramoesian fault. Large-scale NW-wards tilting of Moesia was paralleled by uplift in the East-European/Scythian platform. Most of the present-day shape of the Focşani Depression was achieved during this stage (compare Figs. 4.4 with 4.16C).

Apart from the folding/faulting in the Bend Zone, ascribed to the "Wallachian" phase (sensu Săndulescu, 1984; Hippolyte and Săndulescu, 1996) the deformation pattern of the foreland (Fig. 4.16C) suggests that the Bend orogenic wedge continued to move towards the ESE during this time span.

4.5. CONCLUSIONS

The East-Carpathians foredeep/foreland shows large lateral variations in Neogene sedimentary thicknesses. The general trend is of increasing structural depth towards the orogenic belt from a minimum of 1 km in front of the northern East-Carpathians to almost 13 km in front of the SE-Carpathians Bend, in the Focşani Depression. The present-day Focşani Depression depocenter is NW-SE oriented and is filled with a complete succession of Badenian-Quaternary sediments. From the Badenian onwards, the depocentre of the entire East-Carpathians foredeep has remained in Focşani Depression, although its shape and depth changed through time. During the Badenian the basin extended towards W-NW, beneath the present Carpathians fold and thrust belt. During the Sarmatian, the axis of this depocentre changed to N-S and moved away from the Carpathians. This was accompanied by an important E-wards broadening of the basin. The present shape of the Focşani Depression came into evidence starting with the Pliocene, that is, coeval with the E-wards tilting of its western margin. Consequently, the Focşani Depression was broader prior to the recent times.

Large discrepancies and even contrasting behaviors can be observed between the northern East-Carpathians foreland (East-European/Scythian platform) and the southern one (Moesia) in terms of tectonics and basin subsidence during the contractional and post-contractional stages of the Carpathians. During the Badenian, small and gently increasing W-wards subsidence associated with rather small contractional structures characterized the East-European/Scythian foreland whereas Moesia underwent extension-related subsidence. NE-SW directed extension affected areas that extended from the Focşani Depression to the SE, including Dobrogea, where this deformation was recognized from kinematic indicators (Hippolyte, 2002).

During the Sarmatian, subsidence continued and accelerated in the northern foreland, which was tilted towards the orogen and affected by normal faulting paralleling the strike of the latter in response to the main deformation of the East-Carpathians fold-and-thrust belt. By contrast, the Sarmatian deposition in the southern foreland started later, following an erosional unconformity. In Moesia (except the Focşani Depression), the Sarmatian sequence also thickens towards the orogen but is thinner than on the East-European/Scythian platform (compare for instance Figs. 4.12B with Fig. 4.14A that lie approximately at the same distance relative to the Carpathians thrust front). The main mechanism of tectonic deformation in the southern foreland was strike-slip.

The differences between the two foreland blocks persisted during the Meotian-Pontian, mainly in terms of vertical movements. The East-European/Scythian platform remained stable or experienced uplift whereas Moesia subsided, accompanied by some normal and minor strike-slip faulting. In fact, after the Sarmatian, the subsiding basin was basically limited to the north by the Trotuş fault.

The Pliocene-Quaternary latest stage of basin evolution records the largest difference between the East-European/Scythian platform and Moesia, respectively. Positive movements continued in the former whereas the latter experienced strong tilting and subsidence, together with a rejuvenation of normal and strike-slip faulting. In Moesia, the thickness of the Pliocene-Quaternary sequence approximately equals that of the Sarmatian-Pontian (e.g. Figs. 4.12A and B).

An important inference can be drawn from the subsidence history documented here. The models invoking the SE-wards along-arc detachment of an oceanic slab from the foreland plate being responsible for the anomalously great depth of the Focşani Depression (e.g. Wortel and Spakman, 2000; Sperner et al., 2001) cannot explain the pattern of vertical movements in the foreland, irrespective of whether the detachment would have been progressive or taken place in several sudden events. A consequence of these detachment models would be the migration of the foredeep depocenter along the belt, i.e. from the NW to the SE in our case. Actually, no such change can be documented and the maximum subsiding region remained fixed during the entire foredeep stage (see also Maţenco et al., 2003). Taking into account the recorded evolution of the foreland basin, its subsidence should be rather split into at least two stages: the first is related to the Badenian extension and the second to the flexure in response to orogenic loading. A flexural mechanism is suggested by the regional tilting of the foreland that occurred during and after the Sarmatian major shortening. Localization of the post-Sarmatian subsidence in Moesia and the positive movements recorded in the Carpathians Bend (e.g. Bertotti et al., 2003) suggest that lithospheric buckling occurred as well. The next chapter will address quantitatively the extensional and flexural stages.

CHAPTER 5

POLY-STAGE SUBSIDENCE IN THE CARPATHIANS FOREDEEP: INFERENCES FROM EXTENSIONAL AND PLANFORM FLEXURAL MODELING

5.1. INTRODUCTION

This Chapter addresses the subsidence in the Carpathians foredeep by means of numerical modelling. The emphasis is put on the Focşani Depression as it lies in a peculiar position relative to the orogen and records very great subsidence, in fact the greatest of the entire foredeep (see Chapter 4). The new data interpreted in previous chapters allow for accurate modeling, which has important implications not only for the processes controlling the subsidence of the Focşani Depression but also for the tectonic mechanisms acting in the last stages of continental collision in the Carpathians realm.

As the SE Carpathians foredeep is characterized by two quite different tectonic regimes, NE-SW oriented extension followed by tilting and regional subsidence (Chapter 4), the modeling approach is divided accordingly. The aims of the first modeling computation were to determine the amount of subsidence that can be ascribed to extension, for both syn and post-rift periods, and to estimate the lithospheric characteristics following the extensional event. The second step was the modeling of the flexure of Carpathians foreland due to the present topographical load. The novel approach consists in the planform computation of both the load and lithospheric strength (translated in terms of Elastic Effective Thickness – EET). This approach allows the incorporation of lateral changes in the EETs of the lithospheric domains comprising the Carpathians region. None of the previous modeling studies carried out in the Romanian Carpathians (Royden and Karner, 1984; Royden, 1993; Maţenco et al., 1997a) took into account the lateral rheological variations or the 3D load geometry as they used only 2D computation. Whether the observed subsidence can be ascribed primarily to the 3D distribution of loads/lithospheric strengths is the main objective of this chapter.

It should be mentioned that the modeling studies undertaken for the Romanian Carpathians (cited above) concluded that the load exerted by the present topography of the Carpathians is too small, especially for the depth of Focşani Depression (accounts for maximum 10% of subsidence) and that “hidden loads” are required. The reader is referred to subchapter 2.4 for a review of the models proposed for the foredeep subsidence.

5.2. EXTENSIONAL MODELING

5.2.1. Badenian extensional basin

One of the main results of the basin reconstruction presented in Chapter 4 was the recognition along the eastern margin of the Focşani Depression of a set of NW-SE trending normal faults of Badenian age (Fig. 4.5). Badenian sediments are up to 4 km thick and mainly found along an area elongated in NW-SE direction. The thickness of Badenian sediments rapidly decreases from the Focşani Depression to its eastern margin (on the North Dobrogea promontory). The NW to W-wards prolongation of the basin is unclear since it was buried by younger sediments and the arrival of the Carpathians nappes. Thus, the position and structure of the former western extensional margin of the Focşani Depression remain unknown.

This chapter is mainly based on: Tărăpoancă, M., D. Garcia-Castellanos, G. Bertotti, L. Maţenco, S.A.P.L. Cloetingh and C. Dinu, Role of the 3D distributions of load and lithospheric strength in orogenic arcs: poly-stage subsidence in the Carpathians foredeep; Earth and Planetary Science Letters, in press.

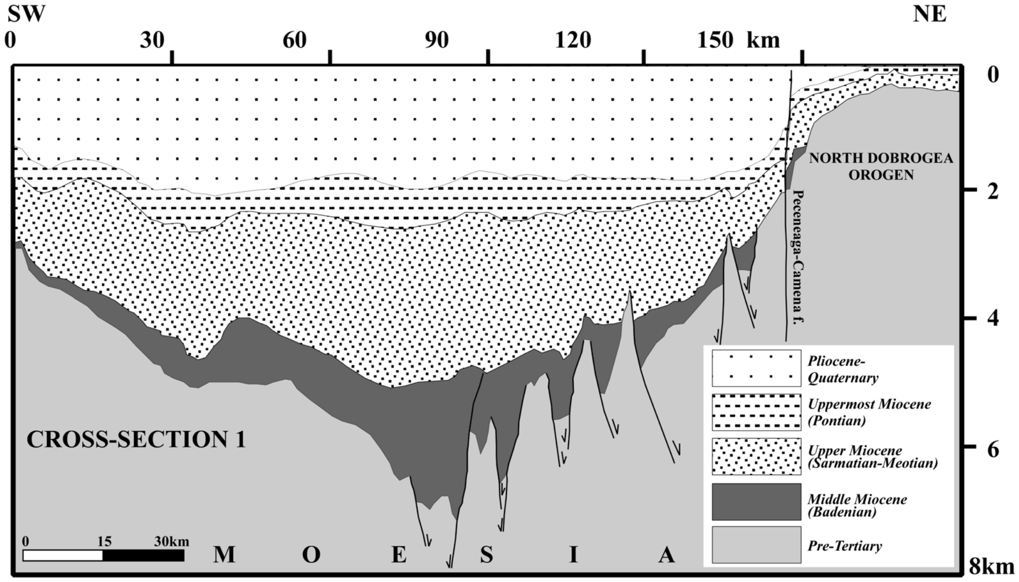


Figure 5.1 *Cross-section CS 1 across the Badenian extensional basins. Location in Figure 4.5.*

The Badenian extension is clearly visible to the SE of Focşani Depression where it produced NW-trending grabens and half-grabens bounded by ENE-striking transfer faults. These grabens are narrow and reach a maximum depth of 2 km (Fig. 5.1; see also Fig. 4.7). The Badenian sequence represents the syn-rift fill of these basins. The Badenian/Sarmatian boundary is represented by an almost regional erosional unconformity. This horizon marks the end of extension and is covered by onlapping Sarmatian strata.

5.2.2. Modeling approach

To model the Badenian extension in the Carpathians Bend foreland, a forward 2D numerical code (Kooi, 1991) was used simulating the kinematic and thermal behaviour of the thinning lithosphere. The lithosphere is divided into a series of vertical boxes with a given width (3 km in our case) and variable thinning factors assigned to each box. The thinning factor is defined here as the ratio between initial to final thickness. Different thinning factors can be assigned to the crust and lithospheric mantle (Royden and Keen, 1980).

Subsidence is computed for the centre of each box. Finite duration of rifting and lateral heat flow is incorporated in the model. From the thinning factors and thermal state of the lithosphere the unloaded subsidence is determined isostatically and the basement subsidence is computed from flexural loading by filling the basin with water and sediments. Sediments fill the space between the imposed paleobathymetry and the basement subsidence. Surface processes, as the erosion of rift shoulders, are not considered. Sediment compaction is taken into account using a simple exponential porosity-depth relation (Sclater and Christie, 1980).

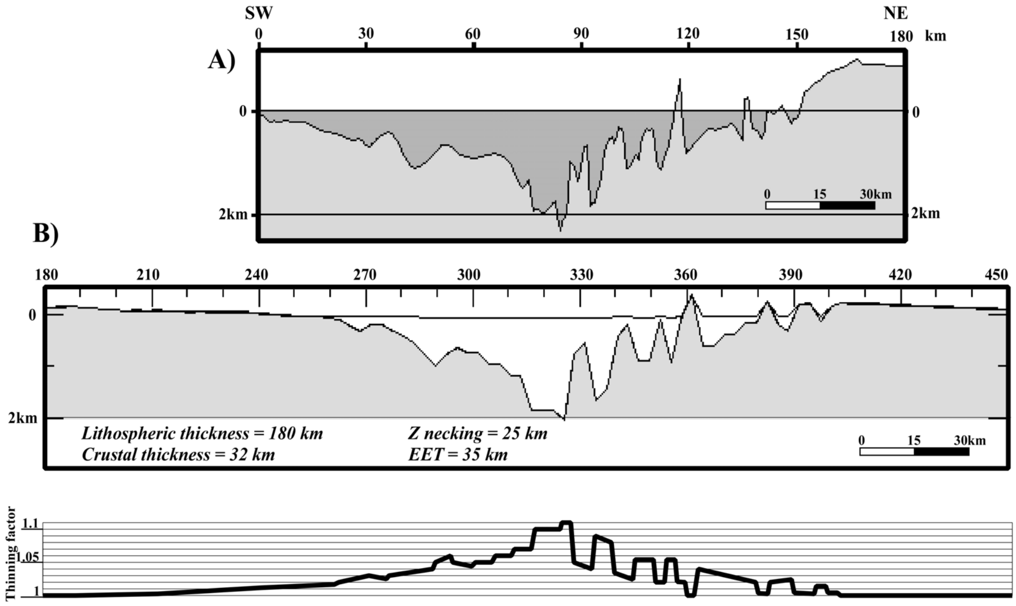


Figure 5.2 *Results and parameters of the extensional modeling for the Badenian basins.*
A) *Cross-section CS 1 (Fig. 5.1) flattened at the Top Badenian horizon.*
B) *Best-fit modeled cross-section and resulting parameters. Also shown are the inferred thinning factors (same for crust and lithospheric mantle).*

5.2.3. Input parameters

The model requires the EET and depth of necking as input values, the former being defined either as a constant or associated with a specific isotherm. Pre-thinning lithospheric and crustal thicknesses represent input values as well. Adopting these parameters, various thinning factors were tested in order to fit the basin geometry. It should be noted that the model results were tested only against the basin stratigraphy and not against features such lithospheric thickness, which have been overprinted by subsequent tectonic events.

The cross-section CS 1 (Fig. 5.1) has been chosen for the 2D modeling because it best images the extensional basins. To obtain the basin architecture prior to thrusting and flexural loading, the Top Badenian horizon was flattened and the underlying features were restored (Fig. 5.2A). Since the Top Badenian horizon does not extend over the North Dobrogea promontory this approach leads to an artifact on the easternmost rift shoulder overestimating its uplift during Badenian. In reality, most of the uplift was produced during the Pliocene (note the thickening of the Pliocene sequence to the west of Peceneaga-Camena fault in Figure 5.1; see also Fig. 4.15B). The modeled section has been made longer than the width of the basin in order to avoid boundary effects.

The foreland of the Carpathians Bend area is characterized by a present day lithospheric thickness ranging between 170-190 km and a crustal thickness between 30-40 km (e.g. Rădulescu, 1988; Horváth, 1993; Fig. 2.5). An unstretched lithospheric thickness of 180 km was chosen as a modeling assumption. Zero paleowater depths were used. This is consistent with the presence of the Badenian shallow water deposits with some evaporitic intercalations on the outer part of the foreland (Saulea et al., 1969; Paraschiv, 1979a). The basin is assumed to be filled with sediments up to sea level and their compaction is taken into account using the values of 0.5 and 0.8 for the surface porosity (as fraction) and characteristic depth constant (km^{-1}) respectively, corresponding roughly to a silty sand lithology. Other parameters used are shown in Table 5.1. The duration of Badenian extension was 3 Myr (Chapter 4).

TEMPERATURE	**Surface**	0C
	Asthenosphere	1333C
DENSITY AT SURFACE CONDITIONS	**Crust**	2.8g/cm3
	Mantle	3.33g/cm3
SEDIMENT GRAIN DENSITY		2.66g/cm3
COEFICIENT OF THERMAL EXPANSION		3.4E-5C
THERMAL DIFFUSIVITY		7.8E-7m2/s

Table 5.1 *Parameters used as input data in the extensional modeling.*

5.2.4. Results

The best-fit model is shown in Figure 5.2B. A good fit is obtained for most of the section (Fig. 5.2A). Only in the westernmost part of it there are some differences, probably related to subsequent deformations along the Intramoesian fault.

The best-fit model was obtained assuming uniform extension, i.e. the thinning factors are the same for crust and lithospheric mantle. Thinning factors are small and even the deepest central basin has a thinning factor of only 1.1 (Figs. 5.2A and B). They decrease more gradually from the centre of the basin towards its western side than towards the eastern one. The best-fit model was obtained using a high EET value of 35 km, an intermediate to deep depth of necking of 25 km and a pre-extension crustal thickness of 32 km.

The thermal anomaly produced by the Badenian rifting is small due to the low thinning factors and probably due to the lateral heat dissipation. Under these circumstances, the post-rift subsidence (post-Badenian) was very low to absent (Fig. 5.3). The maximum predicted post-rift subsidence is less than 100 m.

Changing initial crustal thickness (from 30 to 40 km) has little effect upon the model results. By contrast, the EET and, especially, the depth of necking have strong influence. Assuming the same distribution of the thinning factors as in Figure 5.2 and a crustal thickness of 32 km, basin geometries for four depths of necking and four EET values, respectively, are shown in Figures 5.4A and B. Due to the low magnitudes of the thinning factors, changes in these two parameters affect especially the central and central-western part of the rifted area. It can be observed that the same basin and same amount of thinning are obtained by using a lower EET and a higher depth of necking or vice versa. In other words, an unique solution in terms of EET, depth of necking and lithospheric thinning cannot be obtained using only the basin shape. The preferred values of EET and necking depth characterizing our best-fit model are similar to those obtained by Spadini et al. (1996) for the Mid-Cretaceous rifting in the Western Black Sea. This is consistent with the fact that both basins involved extension of the Moesian platform.

An EET value considered as the depth to the 400^{o}C isotherm has led to an almost identical shape of the basin as the elastic value of 35 km (for a depth of necking of 25 km).

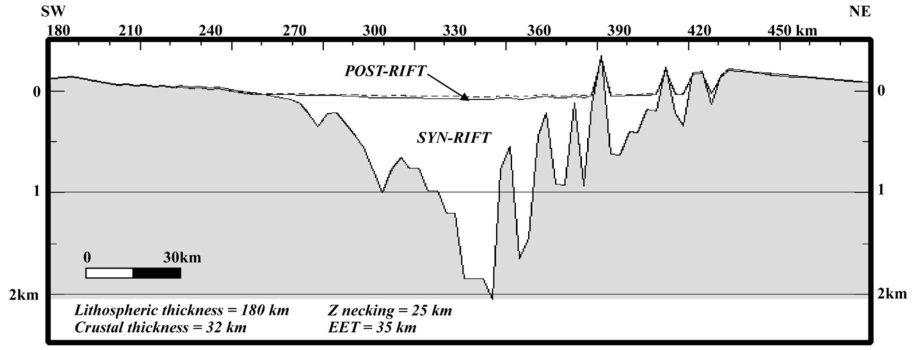


Figure 5.3 *Syn and post-rift subsidence obtained for the best-fit model (shown in Figure 5.2).*

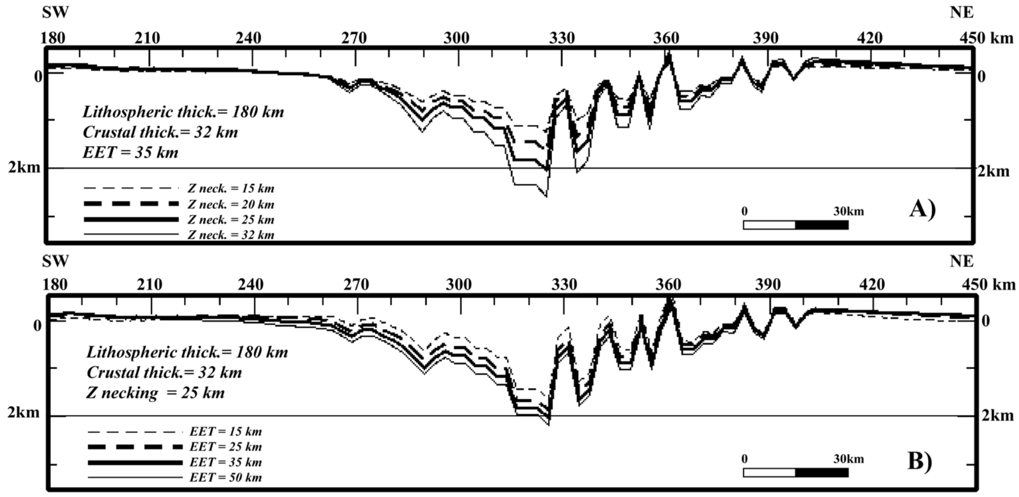


Figure 5.4 *Influence of the variations of depth of necking (**A**) and elastic effective thickness (**B**) upon the basin shape using the same thinning factors as for the best-fit model (shown in Figure 5.2).*

5.2.5. Inferences from extensional modeling

The model predicts almost no post-rift subsidence, which further could be associated with the unconformity at the base of the Sarmatian that is observed on seismic lines (Fig. 4.7). This means that only a very small fraction of the subsidence recorded during Sarmatian times (Fig. 5.1) can be genetically linked to Badenian extension. Creation of the accommodation space for the post-rift sediments will be related in the next part of this chapter to lithospheric flexure induced by the final emplacement of the Carpathians nappes. As a result of the relatively minor Badenian crustal extension, the rift-induced strength decrease of Moesia is expected to be small.

5.3. FLEXURAL MODELING

A topic that has never been addressed by 2D flexural modeling studies in the Romanian Carpathians (Royden and Karner, 1984; Royden, 1993; Maţenco et al., 1997a) is the geographic position of the zone of maximum subsidence, namely the Focşani Depression. The Carpathians foredeep in the front of the Bend Zone is peculiar because the depocentre does not lie under the thrust belt as predicted by simple models. Indeed, isopach maps of the

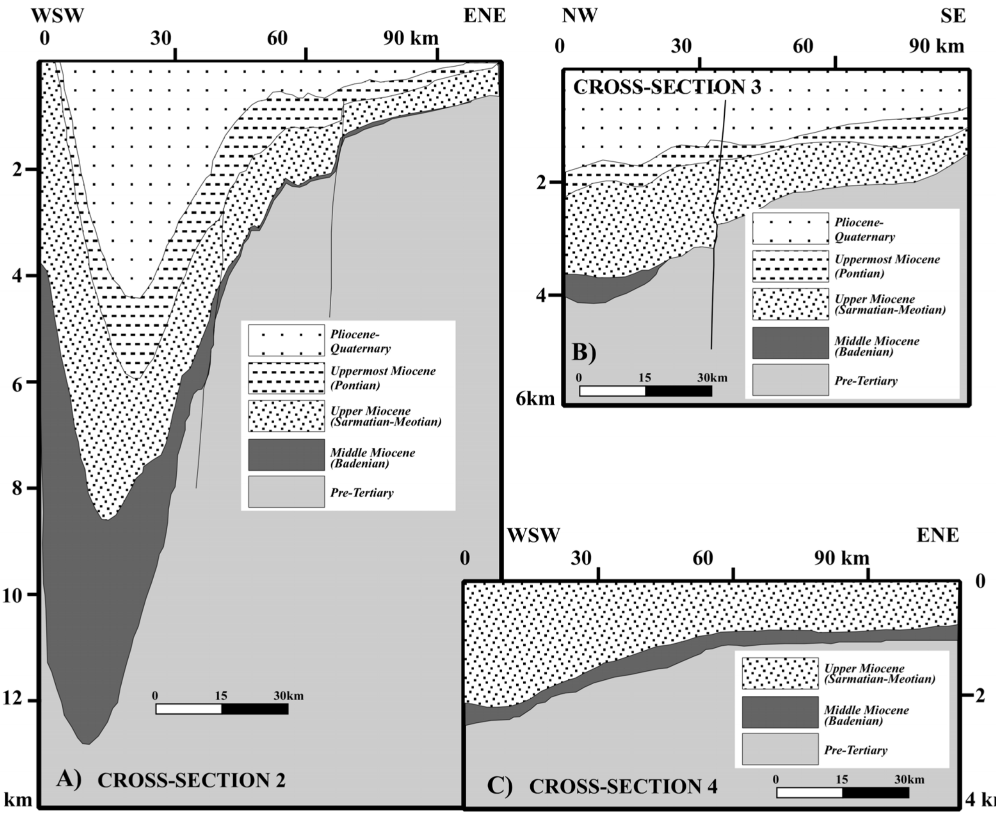


Figure 5.5 *Cross-sections CS 2, 3 and 4 showing the Focşani Depression (**A**), southern foredeep (**B**) and northern foredeep (**C**), respectively. Location in Figure 4.11.*

Sarmatian and younger sequences (Chapter 4, Figs. 4.11 and 4.16) show that the basin floor becomes shallower towards the belt. Here, the possibility is investigated whether this peculiar feature is associated with the presence within the Carpathians domain of lithospheric blocks with significantly different strengths.

Following an initial cross-section experiment, a planform modeling approach was used. Two topics are addressed: (1) the effect of 3D load distribution and lateral variations in lithospheric strengths upon the basin centre position; (2) whether the very large observed subsidence is the effect a 3D loading rather than of a hidden load as inferred by previous 2D experiments.

5.3.1. Sarmatian-Quaternary foredeep basin

The geometry of the Carpathians foredeep has been detailed in Chapters 3 and 4. In this section, only three cross-sections through the foredeep, all associated with the SE-Carpathians Bend are shown (Fig. 5.5). They will be used for comparison with the results of flexural modeling.

Around 8 km sediments accumulated after the Badenian in the Focşani Depression (Fig. 5.5A). In contrast to the eastern margin, which dips gently towards the basin, the western one is steeply E-wards dipping and the Upper Miocene strata crop out with almost vertical dips (Dumitrescu et al., 1970).

To the south and north of the Focşani Depression (Moesia and East-European/Scythian platform, respectively), the foredeep fill thickens in a wedge towards the orogenic belt (Figs. 5.5B and C, respectively). However, in contrast to the East-European/Scythian platform (Fig. 5.5C), the foredeep stage in Moesia comprises sediments as young as Pliocene-Quaternary (Fig. 5.5B). Also, a noteworthy feature observed in Moesia is that the Pliocene-Quaternary sequence has roughly the same thickness as the Sarmatian - Pontian one.

5.3.2. Lateral EET changes and the position of a foredeep depocentre: a 2D approach

Flexural modeling studies of foreland basins show that the width and depth of the basin are dependent on the rigidity of the lithosphere, expressed by its EET, and on the shape and weight of the orogenic load (e.g. Beaumont, 1981; Jordan, 1981). A load imposed on a continuous plate with constant or very gradually varying EET produces a basin centred underneath the load itself. The basin floor gradually shallows on both sides.

The EET of the lithosphere is a function of its crustal and lithospheric thickness and thermal regime. Pre-existing crustal fractures weaken the crust and thus the EET of the entire lithosphere. As the load could be transmitted into the distal foreland, the predicted foredeep position and shape are, however, significantly changed when two adjacent domains with very different strengths form the loaded plate.

To illustrate the effect of lateral changes in plate strength upon basin geometry, the deflection of a 2D thin elastic plate is calculated under the weight of a rectangular load for different EET distributions (Figs. 5.6A and B, using the numerical code of Garcia-Castellanos et al., 1997). In these calculations, the imposed topography after the lithospheric deflection is 2 km in the orogen and the flexural basin is filled with sediments up to the sea level (other parameters are presented in Table 5.2). The issue is relevant because the Carpathians fold-and-thrust belt overlies lithospheric blocks with very different characteristics (see subchapter 2.3). The boundaries between these blocks are often formed by high angle faults (e.g. the Trotuş fault) and are assumed to be sharp.

One can observe that the position of the deepest point of the basin tends to shift away from directly below the load towards the weaker domain ("foreland plate"). The shape of the basin also differs from the one predicted for a constant EET (Fig. 5.6A). The trend can continue and the site of maximum deflection can be laterally offset from the load itself when the difference in EETs becomes significant (Fig. 5.6B). From this point, the base foredeep becomes rapidly shallower towards the orogenic load. The depocentre lies in front of the load block and has a steep external flank like that observed in the Focşani Depression.

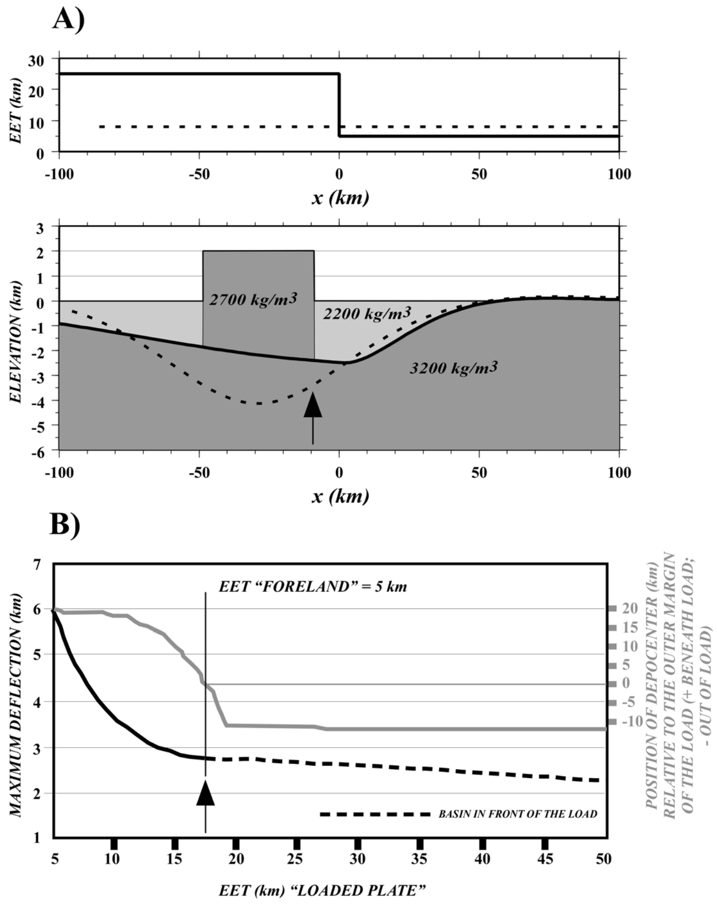


Figure 5.6 *(**A**) Flexural deflection profiles obtained for two different EET distributions using the same topographic load. A constant EET value predicts maximum deflection under the load (dotted line), whereas an abrupt decrease in the foreland EET can induce maximum deflection in the weaker foreland (bold line).(**B**) Maximum deflection as a function of the strength of the loaded plate (black line) and the position of depocentre relative to the load (gray line).*

5.3.3. Planform modeling approach

To investigate the mechanism of subsidence in the Focşani Depression, a flexural model was used that calculates the deflection corresponding to a laterally variable load distribution (Garcia-Castellanos, 2002). The load is calculated from the observed topography in the Carpathians region and the deflection is filled with constant density material. A detailed description of the equations governing this flexural plate can be found in van Wees and Cloetingh (1994). The grid covering the Carpathians region is composed of 10,000 points on the base of which the computations of both load and deflection are performed. The final deflection pattern predicted by such a model, and in particular the localization of maximum deflection, is dependent not only on the topographic load distribution, but also on the elastic thickness distribution EET(x,y).

5.3.4. Input data

The Focşani Depression lies on the northern part of the eastern Moesian platform and is surrounded by other lithospheric blocks with quite different tectonic histories and, therefore, different strengths. The structure of the Carpathians domain is divided into five major lithospheric blocks (Fig. 5.7A): East-European platform (inclusive of Scythian platform and North Dobrogea orogen), eastern and western Moesian platforms (the latter includes the South-Carpathians/Getic Depression as well), East-Carpathians and Transylvanian (including the Apuseni Mts. and the easternmost part of the Pannonian basin).

DENSITY	**Load**	2700 kg/m3
	Sediments	2200 kg/m3
	Asthenosphere	3200 kg/m3

Table 5.2 *Densities used in flexural modeling.*

The South-Carpathians and Getic Depression are "incorporated" in the western Moesian block in order to avoid numerical instabilities that might result from the incorporation of small domains insufficiently covered by the grid points. In the modeling approach, the lithospheric blocks behave elastically and the EET is assumed to be constant. Major wrench faults represent the boundaries between the different foreland blocks. The Carpathians are separated from the foreland by the Pericarpathian fault, which is the most external major thrust (e.g. Săndulescu, 1988). The boundaries between the Transylvanian block and the East- and South-Carpathians could be represented by a backthrust (Huismans et al., 1997; Sanders, 1998) and by a dextral shear zone (e.g. Linzer et al., 1996), respectively.

The densities of the asthenosphere, the load and the material filling the deflection are the same as those given in Table 5.2. The EETs are required as input values. The geophysical characteristics of the defined blocks, comprising the Carpathians realm, are fairly well known (Table 5.3), but while the tectonic differences between these blocks are obvious, a translation to EET values is far from straightforward (e.g. Burov and Diament, 1995). Therefore, sets of EET were chosen in such a way that: (1) the relative order given in Table 5.3 is respected; and (2) the range of absolute values is closer to the rheological estimates (Lankreijer et al., 1997) than to published flexural ones (Maţenco et al., 1997a). Except for Royden's estimate (Royden, 1993), the EET values obtained from flexural studies of the Carpathians/foreland system are much lower (at least half) than those inferred from rheological studies (Table 5.3). EET values lower than 15 km obtained by Maţenco et al. (1997) appear to be unrealistic for some lithospheric domains, such as the East-European craton (see the Comments in Table 5.3). On the other hand, for eastern Moesia both flexural and rheological estimates did not consider the effects of the Badenian extension and major crustal faults, or structural heterogeneities sensu Burov et al. (1998). The EET of such pre-weakened lithosphere could have been further affected by strong bending (e.g. Waschbusch and Royden, 1992; Burov and Diament, 1995) especially in the Focşani area.

It is assumed that development of the foredeep commenced at the beginning of the Sarmatian. For this time the elevation of the entire model is taken as zero metres. This is justified as the top Badenian/base Sarmatian is formed in (1) foreland by either mainly evaporitic (on East-European/Scythian platform), or fluvial sediments (parts of western Moesian platform) or corresponds to a hiatus (parts of western Moesian platform and eastern

LITHOSPHERIC DOMAIN	CRUST (km)	LITHOSPH. (km)	EET ESTIMATES (km)	COMMENTS
E-Europ. (including Scythian and N-Dobr orogen)	> 40	> 190	14 (plate boundary forces acting) [1] 29-32/38-79 (wet/dry rheology) [2] 40 (subsurface loads) [3]	oldest, coldest and strongest domain
Eastern Moesian platform	30-40	150-190	12 (plate boundary forces acting) [1] 24/39 (wet/dry rheology) [2]	weakest domain due to rifting + crustal scale shearing + bending
Western Moesian platform (including South Carpathians)	30-45	150-180	<15 (plate boundary forces acting) [1] 24-72/29-84 (wet/dry rheology) [2]	old, cold, strong domain
Transylvanian (incl. Apuseni Mts.)	27-35	75-110	12-25/19-32 (wet/dry rheology) [2] 20 [4]	cold, intermediate strength
East Carpathians	40-45	120-170	~ the same as E-Europ. domain [1] 19/32 (wet/dry rheology) [2]	weaker than E-Europ. platform due to the bending + Neogene volcanism
Crustal and lithospheric thick.:			Radulescu, 1988; Horvath, 1993; Nemcok et al, 1998.	
EET estimates:			[1] Matenco et al, 1997 (flexure); [2] Lankreijer et al, 1997 (rheology); [3] Royden, 1993 (flexure); [4] Sanders, 1998 (flexure).	

Table 5.3 *Crustal/lithospheric thicknesses and previous EET estimates of the domains comprising the Carpathians region.*

Moesian platform); (2) by salt and shallow-water sediments in Transylvanian basin; (3) by evaporitic and shallow-water deposits in the Outer Carpathians (for 1, see the references from Chapters 3 and 4; for 2, see for instance Huismans et al., 1997; Ciulavu, 1999; for 3, see for instance Ştefănescu et al., 2000). As far as the load-causing subsidence is concerned, the present day topography of the area was used (Fig. 5.7B). The load provided by the water of the Black Sea was neglected. The topography in the Balkans was not considered because the last major deformations there pre-date the Sarmatian (Doglioni et al., 1996; Georgiev et al., 2001). For the same reason (e.g. Săndulescu, 1984) the topography of the Apuseni Mts. is assumed to represent an average for the Transylvanian basin (400 m).

5.3.5. Results

5.3.5.1. Topographic load

If we assign realistic EET values to the domains comprising the study region, a local basin can be predicted to lie in front of the Carpathians Bend, in the area of the Focşani Depression (Fig. 5.8A). Although the overall shape and the eastern margin of this basin are in satisfactory agreement with those of the actual Focşani Depression, the predicted depth is shallower (3.3 km; Fig. 5.8B). The deflection pattern shown in Figure 5.8A is influenced by the strengths of the East-European and western Moesian platforms (EETs of 37 km and 30 km, respectively), the intermediate Transylvanian and East-Carpathians domains (EETs of 26 km and 29 km, respectively) and a weak eastern Moesia (EET = 10 km). The latter causes the localization of the basin along the margin of the load (offset to the east of the orogen), as shown in the previous 2D experiment (Fig. 5.6).

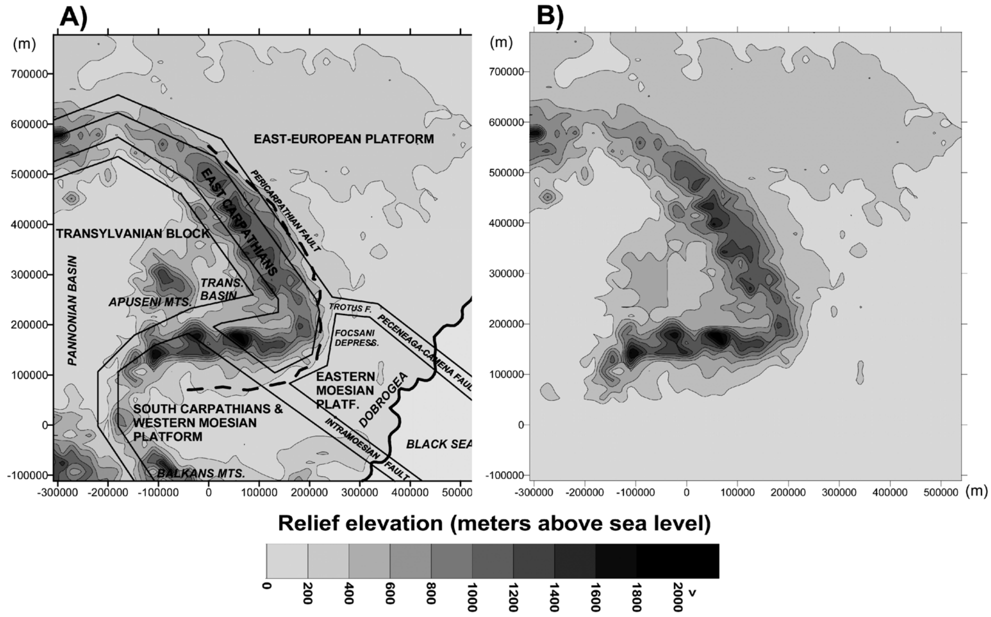


Figure 5.7 *(**A**) Lithospheric domains used in the 3D flexural modeling. Boundaries between them are represented by major faults; (**B**) Topographic load used. Also show is the elevation scale (details in text).*

In front of the East-Carpathians, the predicted depth of the foredeep is between 2.3 and 2.9 km, gradually decreasing E-wards to 1–1.2 km. Close to the Trotuş fault, the predicted deflection fits the base foredeep in the west but overestimates it in the distal parts (Fig. 5.8B), possibly due to the Sarmatian and younger tectonics associated with this fault. In the central-northern part of the East-European platform, the actual base foredeep at the contact with the orogen has depths ranging between 0.8 and 2.5 km, decreasing from SE towards NW along the belt (Fig. 4.11). Farther to the NW, at the transition from the East to the West Carpathians, the base Sarmatian reaches again 1.8–2 km (Fig. 4.10). Thus, the predicted results fit reasonably the actual foredeep geometry, the only discrepancies being related to along-strike depth changes and a shallower position of the basin in the northern East-Carpathians.

To the SE of the Focşani Depression, the modeled basin shown in cross-section CS 3 (Fig. 5.8B) displays depths decreasing from ~2.7 km in the NW to ~1 km in the SE. Although the predicted deflection is shallower than the actual one, it has the same dip. Changes in the foredeep depth and dip occur along crustal foreland faults that represent the boundaries of eastern Moesia and are partly in agreement with offset documented on seismic lines (Chapter 4).

The deflection predicted for western Moesia (South-Carpathians foredeep) is ~1.9 km at the contact with the Pericarpathian fault decreasing W-wards to 1.6 km (Fig. 5.8A). The modeled basin shallows towards the south to ~0.3 km. The map of actual western Moesia presented in Chapter 3 (Fig. 3.2A) shows that the foredeep depths vary between 2 and 3 km along the front of the Subcarpathian nappe and decrease to a few hundreds of metres in a S-SW direction. The fit between predicted and actual deflections could be considered satisfactory for most of the eastern part of western Moesia (compare Figures 5.8B with 3.2A). However, in the western part the predicted basin is significantly shallower than the actual one.

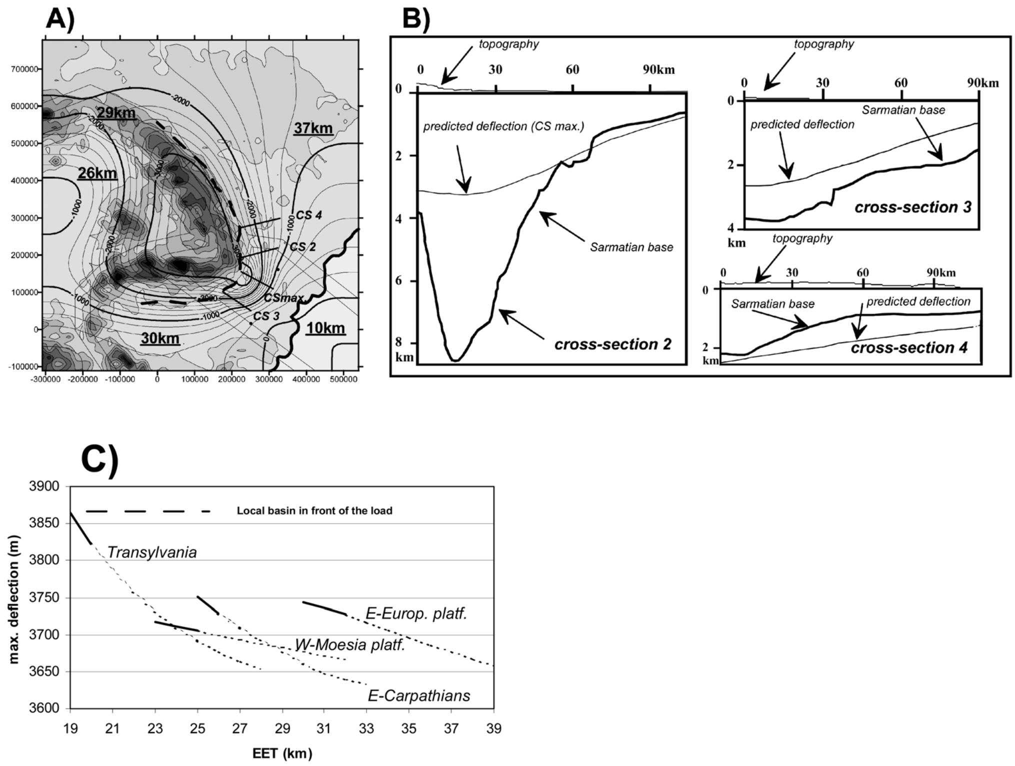


Figure 5.8 *(**A**) Predicted deflection for the topographic load. Underlined numbers represent the EET of each lithospheric block (constant inside the block, linearly interpolated in the boundary polygons). The contours are in meters. (**B**) Cross-sections showing the actual and predicted basin. Location of cross-sections in **A**). CS 2 and CS max are cross-sections through the actual and predicted depocentres. Modeling predicts a depocentre with ~30 km SSW than observed. (**C**) Maximum of deflection as a function of the lithospheric domain strength. The "reference EET values" are those from **A**). The theoretical results were checked for numerical instabilities that might occur due to smaller EET in Eastern Moesia. Note that the changes in EETs are consistent with maximum deflection, i.e. the stronger a domain the shallower the deflection.*

The maximum deflection of ~3.7 km is predicted for the Carpathians hinterland, in the SE corner of the Transylvanian basin (Fig. 5.8A). From this depocentre, the predicted Transylvanian basin shallows to 2.5-2.6 km in a WNW direction. The model satisfactorily fits the actual depth range and shape of the basin (geological cross-sections in e.g. Huismans et al., 1997 and references therein) and also supports the results of a previous model, which linked the post-Badenian salt subsidence in Transylvanian basin to a flexural mechanism (Sanders, 1998).

5.3.5.2. Misfits between predicted and actual basin geometry: possible causes

The use of a forward modeling technique implies that the set of best fitting EET values is not unique. In order to concentrate subsidence out of the orogen, for example, a weak eastern Moesian platform is clearly required, with an EET value less than 12 km. The EET value of the western Moesian block, in contrast, is poorly limited between 24 and 35 km because we cannot quantify at this stage the changes in strength due to the different tectonic events discussed in Chapter 3. The high value for the East-European platform is the most reliable estimate because it is consistent with the thickest, oldest and coldest lithosphere (Table 5.3).

The present topography gives a satisfactory explanation for most of the flexural subsidence of the Carpathians foreland and hinterland. Our modeling approach has assumed an elastic behaviour and the EETs assigned to the different lithospheric domains of the Carpathians region are generally consistent with rheological estimates (Lankreijer et al., 1997). However, our predictions notably underestimate the actual basin depths in the Focşani Depression including eastern Moesia, and the NW part of western Moesia. These misfits persisted even when a viscoelastic rheology was used instead of elastic one.

Some correlations can be drawn between the areas where these misfits are recorded and the overall evolution in terms of tectonics and vertical movements: (1) both occur in Moesia in areas that are associated with orogenic bends; (2) both areas were affected by extension prior to their flexural subsidence; (3) both areas record large subsidence that continued after the emplacement of the Subcarpathian nappe; (4) both areas are affected by young to still active tectonic deformation, post-dating the last major shortening event; (5) both areas are associated with important Pliocene-Quaternary positive movements in the adjacent segments of the orogen, particularly the SE Carpathians Bend.

Whereas crustal extension prior to the flexural stage led to minor to no post-rift subsidence in eastern Moesia (subchapter 5.2), this may not be the case for the NW part of Moesia (see Chapter 3), where the amount of extension may have been larger. Correspondingly, thermal relaxation of the lithosphere may have contributed to the subsidence of the NW Moesia (an example of the effect of rifting prior to flexural loading can be found in Deségaulx et al., 1991).

One of the main problems remains the important tilting documented in eastern Moesia during the Pliocene-Quaternary in the absence of coeval accommodating structures (see Chapter 4). Let us assume that no subsidence occurred during the Pliocene-Quaternary in the Focşani Depression and to the S-SE of it. By comparing the results of planform flexural modeling (Fig. 5.8A) with the configuration of the Focşani Depression restored to the base of the Pliocene, it can be observed that the resulting Sarmatian-Pontian basin averages the predicted deflection pattern (Fig. 5.9). Remember that the Focşani Depression extended farther to the west before the Pliocene, and consequently, the predicted local basin from the Figure 5.8A also would satisfactorily simulate the location of the pre-Pliocene "actual" basin. This means that if no subsidence had occurred during Pliocene-Quaternary times, the Carpathians Bend foredeep (underlain by eastern Moesia) would be a "normal" basin whose depth reflects well the elevation of the adjacent mountain belt.

What could be the cause of the Pliocene-Quaternary subsidence in Moesia, which is largest in the Focşani Depression? As it is unlikely that the Pliocene-Quaternary subsiding centres of the Focşani Depression and NW Moesia are related to separate oceanic or delaminated continental mantle slabs, still attached to the lithosphere (in addition to the arguments against from Chapter 4), lithospheric buckling sensu Cloetingh et al. (1999) is proposed as the main cause for subsidence development. This mechanism has been shown to be consistent with the pattern of vertical movements documented in the Carpathians realm and with the direction of coeval regional compressional stresses (Bertotti et al., 2003). The older extensional strain that is recorded all along the northern margin of Moesia, as well as the presence of major active structural heterogeneities, particularly in its eastern part, may have contributed to weakening of the lithosphere, thus favouring its buckling. This may be the

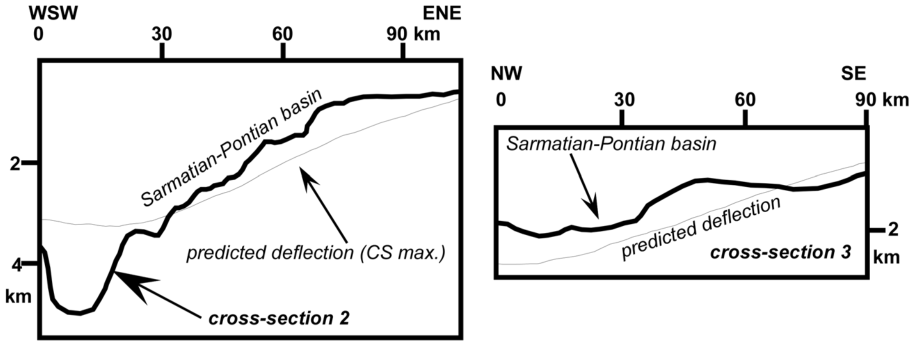


Figure 5.9 *Cross sections 2 and 3 restored at the base Pliocene (the Pliocene-Quaternary sequence is removed from the corresponding cross-sections shown in Figure 5.5). Discussion in text.*

reason why Pliocene-Quaternary subsidence is only recorded in Moesia whereas the strong East-European/Scythian platform behaved as a stable-to-uplifting region. However, at this stage we were not able to incorporate in a model both flexural and buckling processes. Consequently, we cannot use numerical modeling to provide an estimate for the amount of buckling-induced subsidence.

As planform flexural modeling was used in a “static” manner, i.e. instantaneous loading by present-day topography, it is possible that we underestimated the tectonic load due to the highly arcuate shape of the orogen and the diacronous age of topography generation (according to Sanders, 1998, and Sanders et al., 1999, it is mostly Sarmatian in the central-northern East-Carpathians and South-Carpathians and beginning of the Pliocene in the SE Carpathians Bend). This may be the case of the SE Carpathians Bend, which is the region where the last contractional out-of-sequence structures (although minor by comparison with the Sarmatian shortening) are documented (see Chapter 2). However, in the SE Carpathians Bend, Sarmatian thrusting, although obvious (e.g. Mațenco and Bertotti, 2000 and references therein), created no topography, which contrasts with the other segments of the belt. That is, the topography in the SE Carpathians Bend apparently reflects only the last tectonic loading event.

In this region, a supplementary load was added simulating the effect of the Subcarpathian nappe. Figure 5.10A shows the contour of this additional load required to fit the data in the Bend zone. This load can be expressed in various ways, but we chose to apply an additional topography to the Bend orogenic region, with the same density distribution as explained before. Two situations gave significant modeled results, from the variety of loads applied, which were represented by an extra-topography of 500 m and 800 m high, respectively. The main effect for both scenarios is to deepen the satellite basin in front of the Carpathians Bend. The former value was chosen because it could be seen as the average difference in elevation between the East/South-Carpathians and the Bend region. The latter was found to fit the depth in Focşani Depression, but not the geometry (see the next section).

5.3.5.3. The effects of additional load and intraplate stresses

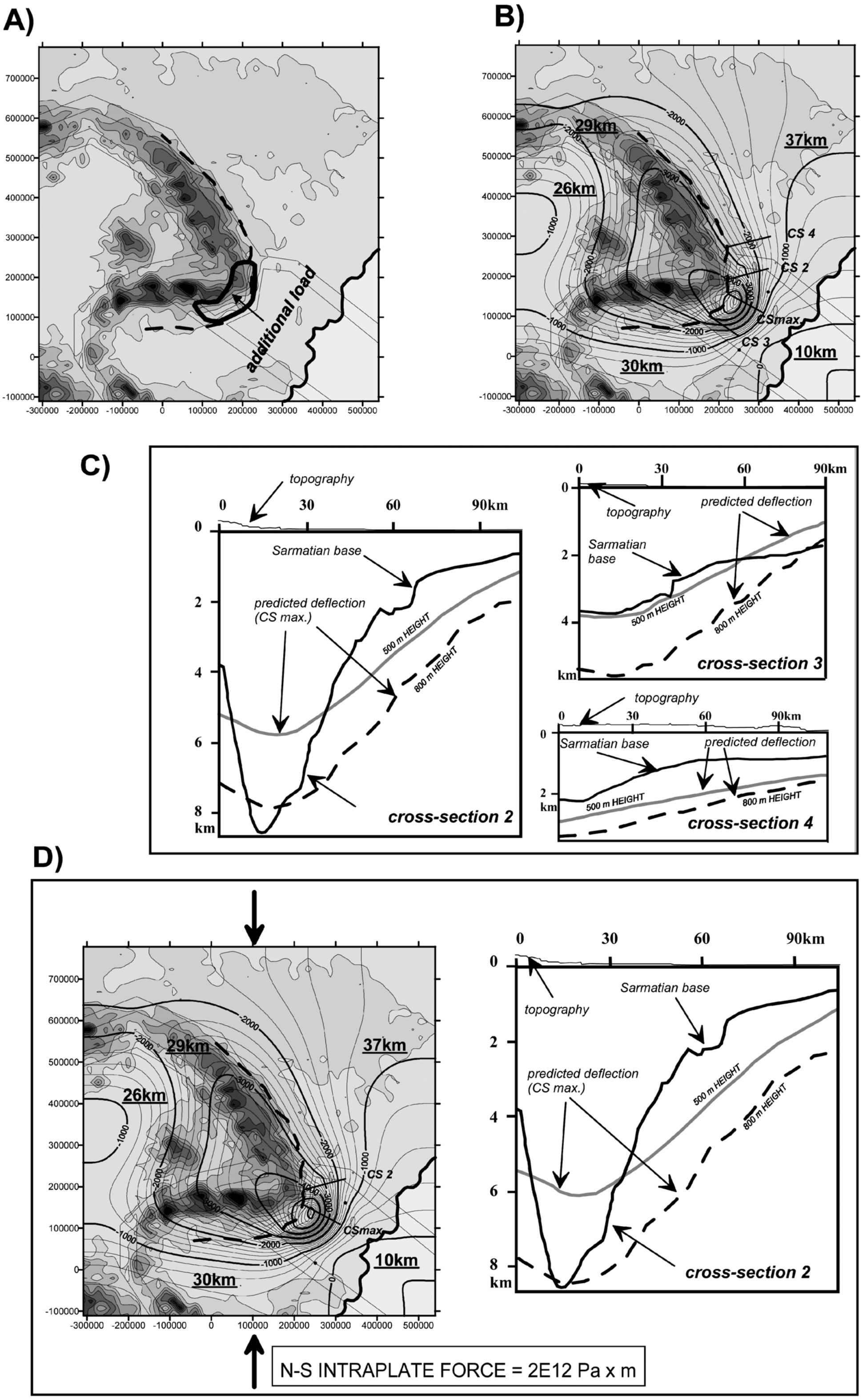
The modeled deflection pattern is shown in Figure 5.10B for the case of present plus 500 m extra-topography. The rheology (pure elastic) and EETs are the same as before. With this extra-topography, the basin predicted in front of the Carpathians Bend is 5.7 km-deep accounting for ~70% of the observed value, but the deflection pattern represents the best possible average of the basin shape (Fig. 5.10C), particularly south of the Focşani Depression (Fig. 5.10C, cross-section 3). The difference observed in the SE part of the cross section 3 is probably related to post-Badenian transtension (Chapter 4). The position of the depocentre fits satisfactory the actual Focşani Depression. By comparing with the results obtained using only the present relief (Fig. 5.8A), it appears that the additional load in the SE Carpathians Bend region (Fig. 5.10B) increases the deflection only in the Focşani Depression, eastern Moesian platform and in SE-most corner of the Transylvanian basin. On the southernmost part of the East-European platform (Fig. 5.10C, cross section 4), the difference between the predicted and the actual basin depth is roughly the same as that obtained without an additional load (Fig. 5.8B, cross-section 4).

The second case corresponds to 800m extra-topography added to the present topography. In this case, the basin predicted in the Focşani Depression reaches a depth of ~8km, which accounts for >90% of the actual value (Fig. 5.10C, cross-section 2). The deflection pattern predicts, however, a basin with a larger wavelength than observed. Also, this 800 m extra-topography significantly overestimates the depth of the eastern Moesia and the East-European platform (Fig. 5.10C, cross-sections 3 and 4, respectively).

Figure 5.10D shows the effect of adding N-S compressional intraplate stresses upon the deflection pattern (the model in the map view is loaded by the present topography plus the 500 m extra-topography in the SE Carpathians Bend). The direction and type of intraplate stresses approximates the Pliocene-Recent ones (Hippolyte and Săndulescu, 1996; Bada et al., 1998; Maţenco and Bertotti, 2000). The predicted deflection in the Focşani Depression reaches 6.3 km, accounting for ~75 % of the amount of post-Badenian subsidence of the actual basin. Moreover, the modeled depocentre coincides closely with the Focşani Depression. In the eastern part of western Moesia, the predicted depth contours become WNW-ESE oriented and the depth increases by a few hundred of metres relative to the model that uses only the present topography (compare Figs. 5.10D with 5.8A). It approximates better the actual subsidence of the eastern part of western Moesia (Fig. 3.2A).

In the second case (800m extra-topography), compressional intraplate stresses would produce a depocentre with equal depth and very close to the location of Focşani basin (Fig. 5.10D, dash line in the cross-section). However, similarly with the scenario without intraplate stresses, the predicted wavelength is larger than the observed one. Modifying the magnitude of the intraplate stresses induced vertical variations in the order of a few hundreds of metres in the depocentral area

Figure 5.10 (next page) *(**A**) Location of the additional load added to the topography. Discussion in text. (**B**) Predicted deflection for the actual topography plus the 500 m-high additional load. The underlined numbers represent the EET of each lithospheric block (constant inside the block, linearly interpolated in the boundary polygons). The contours are in meters. (**C**) Cross-sections showing the actual and predicted basin. Location of cross-sections in **B**). Solid gray line represents predicted deflection for a 500 m-high additional load, whereas dash line refers to an 800 m-high additional load. (**D**) Predicted deflection (map view) for the actual topography plus the 500 m-high additional load and with the contribution of a N-S intraplate compressive force of 2 x 10^{12} Pa x m. In cross-section, solid and dash lines have the same meaning as in **C**).*


A)
B)
additional load
29km
37km
26km
30km
10km
CS 4
CS 2
CSmax
CS 3
C)
topography
Sarmatian base
predicted deflection
(CS max.)
500 m HEIGHT
800 m HEIGHT
cross-section 2
cross-section 3
cross-section 4
D)
N-S INTRAPLATE FORCE = 2E12 Pa x m

5.3.6. Discussion

Taking into account for the development of the SE Carpathians Bend foredeep a Badenian phase of extensional subsidence, followed by thrust-loaded flexural subsidence, ~60% of the depth of the Focşani Depression can be explained using only the present topography as the load for the flexural period. The 3D topography predicts larger flexural subsidence of the foredeep than the previous 2D computations (e.g. Royden, 1993). The most important result is that the 3D distribution of the load and lateral variations in lithospheric strength, particularly the presence of a weak domain in front of the Bend region, determine the localization of a depocentre in front of the orogen, which has a shape that fits well the geometry of the Focşani Depression.

An additional topographic load in the SE Carpathians Bend, together with a contribution from compressional intraplate stresses, can predict most of the flexural subsidence in Focşani Depression. The two case scenarios (500 and 800 m extra-topography in the SE Carpathians Bend, load contour in Fig. 5.10A) can either average the basin deflection and shape, the obtained values being still lower than the observed ones (with ~2 km in the Focşani Depression), or, in the latter case, we can entirely fit the maximum deflection, but not the wavelength.

Because many papers proposed that the subsidence of the Focşani Depression is related to the presence of an oceanic or delaminated continental mantle slab still attached to the lithosphere beneath the SE Carpathians Bend (see subchapter 2.4), we computed the deflection pattern produced by the present topography plus such a slab. The surface projection of the additional load shown in Figure 5.9A is roughly the same as that of the invoked slab (Gîrbacea and Frisch, 1998). We found that the 800 m extra-topography corresponds to an 86 km-long slab attached at the base of the underthrusted continental lithosphere with a density of 25 kg/m^3 representing the density contrast between the sinking slab and asthenosphere. Previous proposed lengths of the invoked slab beneath SE Carpathians Bend are 150 km (Chalot-Prat and Gîrbacea, 2000), 200 km (Gvirtzman, 2002) or 300 km (Wortel and Spakman, 2000). Although the precise value of the density contrast is difficult to constrain, it appears that such a long slab, no matter its origin, produces a too deep basin, also with too large wavelength.

Consequently, three successive distinct processes are proposed for the subsidence in Carpathians foreland basin: extension (which affected only the northern margin of Moesia), flexure and finally buckling. The lithospheric buckling probably started at the beginning of the Pliocene (Bertotti et al., 2003) and is expected to be responsible for 2 km subsidence in Focşani Depression. The resulting overall rate of 0.4 mm/yr for the Pliocene-Quaternary buckling-related subsidence appears reasonable. In the other foredeep region with large subsidence, i.e. the NW-most part of Moesia, the results of the planform modeling show probably only the contribution of the flexure to the overall subsidence. Here, the remaining 2 km of subsidence could be seen as resulting from thermal cooling that followed the Burdigalian (and Paleogene?) extension and the young lithospheric buckling. Thus, lithospheric buckling appears as the tectonic mechanism that was active in the last stages of the convergence and controlled the pattern of vertical movements within the entire Carpathians/Pannonian region (see also Horváth and Cloetingh, 1996).

Dealing with pure elastic plates loaded instantaneously, the modeling has considered neither changes in the lithospheric strength that may have occurred in time due to thermal events, faulting, bending (Waschbusch and Royden, 1992; Burov and Diament, 1995; Burov et al., 1998) and thickening/thinning of the sedimentary cover (Lavier and Steckler, 1997), nor those that affected the topography (load). Given all these factors involved in the temporal changes of the lithospheric strength, a planform kinematic flexural modeling should be the next step towards a better explanation of the Carpathians foredeep subsidence. A full integration of the rheological behavior is also imposed.

5.4. CONCLUSIONS

The subsidence of the unusually deep SE Carpathians foredeep basin is presumably mainly the effect of pre-orogenic extension followed by topographic load-induced planform flexure. NW-SE trending extensional basins are found to the south of Focşani Depression, whose eastern margin underwent also normal faulting. Extensional modeling for these basins reveals a uniform crust/mantle lithosphere thinning of maximum 1.11. This modeling also predicts that after the Badenian extension, very small or no post-rift subsidence occurred. This is in accordance with a regional unconformity observed at the Badenian/Sarmatian boundary.

For the second subsidence stage (post-Badenian), a flexural model was applied that takes into account the 3D distribution of a topographic load and lateral variations in the lithospheric strength. Assigning realistic strengths to the domains comprising the Carpathians region, a basin is predicted to exist in front of the Carpathians Bend because the weakened foreland (due to extension and crustal active faults) localizes the lithospheric deflection. Since the deepest modeled basin is 3.3 km that represents only 40 % of the observed basin depth, an additional load is required in the Carpathians Bend, corresponding to a supplementary topography with an elevation of 500 m. Depending on the magnitude of intraplate stresses, a basin could develop at the location and with the shape of the Focşani Depression and 75 % of its depth could be accounted for.

It is possible that a planform kinematic model (rather than an instantaneous-loading one) that also incorporates temporal variations in the lithospheric strength, could fully explain the observed subsidence without any additional load. Alternatively, the last stage of subsidence (Pliocene-Quaternary), which can be documented only in Moesia, may be the result of lithospheric buckling.

An alternative has been provided to models invoking a slab pull force for the large subsidence observed in the SE Carpathians foredeep/foreland, particularly in the Focşani Depression. This contribution demonstrates (1) the importance of the previously neglected extension in the eastern Moesian foreland and, (2) that the 3D distributions of the topography and lithospheric strength, plus the effects of compressive intraplate stresses can lead to major subsidence in front of the orogenic load. This 3D effect could be particularly important for foredeeps associated with highly arcuate orogenic belts. This can be clearly observed in the Carpathians region where basins associated with the most arcuate parts of the orogen, no matter whether they lie on the convex or concave side, experienced the largest subsidence along the entire foredeep system.

CHAPTER 6

SUMMARY AND GENERAL CONCLUSIONS

6.1. INTRODUCTION

To fully understand the evolution of the Carpathians foredeep/foreland basin, its integration into the regional tectonic framework is essential. A review of the pattern of vertical movements recorded in the Carpathians foredeep/foreland together with the tectonic evolution of the Carpathians orogen is presented in the following for several Tertiary time frames. The resulting schematic maps integrate the findings of Chapters 3 and 4 for the foreland basin with the development of the orogen and the Transylvanian basin.

Large variations in terms of vertical movements along the Carpathians foredeep/foreland are associated with variable amounts of exhumation taking place at different times within the orogen (Fig. 4.2). A correlation will be made between the subsidence/uplift in foreland and exhumation in the Carpathians orogen, highlighting the importance of rheological variations in the lower plate on the overall tectonic deformations.

6.2. PALEOGENE

Following the Late Cretaceous shortening ("Laramian phase" sensu Săndulescu, 1984, 1988) the South-Carpathians was located S to SSW of Moesia (Fig. 6.1A; also Csontos, 1995; Schmid et al., 1998). South-Carpathians underwent an important extensional collapse to the end of its collision with Moesia, the restored extensional direction being roughly NNW-SSE (Willingshofer, 2000; Willingshofer et al., 2001). In this context, it should be noted that Fügenschuh and Schmid (submitted) proposed that only erosion-driven cooling took place in the orogen units during the Late Cretaceous-Paleocene. Subsequently, the South-Carpathians units moved NE-wards and rotated clockwise around Moesian corner during the Eocene-Oligocene, accompanied by core-complex formation and orogen-parallel extension (Fig. 6.1A; Schmid et al., 1998; Mațenco and Schmid, 1999; Fügenschuh and Schmid, submitted). Eocene extension is also documented in the southern part of East-Carpathians (Săndulescu, 1992).

The structural configuration of the northern margin of Moesia (i.e. below the present Getic Depression) is poorly known due to the great thickness of the Paleogene sequence: these thick Paleogene deposits, which crop out at the contact with the Mesozoic nappe pile (e.g. Jipa, 1980, 1984), are interpreted either as a foredeep (e.g. Săndulescu, 1984) or as the infill of an extensional/transtensional basin (Mațenco et al., 1997b). The latter hypothesis seems more appropriate in the context of the overall extension in neighbouring areas (Fig. 6.1A). Coeval with the orogen-parallel extension in South-Carpathians, the Balkans were thrusted during the Eocene on top of the southern margin of Moesia (Doglioni et al., 1996; Georgiev et al., 2001).

Following the orogen-parallel extension of the South-Carpathians, dextral wrenching accommodated its continuous displacement around the Moesian corner during late Eocene-Oligocene times (Fig. 6.1B; Ratschbacher et al., 1993; Fügenschuh and Schmid, submitted). Dextral displacements occurred inside both the orogen (along the Cerna-Timok fault system) and in westernmost Moesia. No other active structures could be documented in the Carpathians foreland.

The Eocene represents the onset of positive vertical movements in western Moesia as evidenced by the development of a regional erosional unconformity (Fig. 6.1A). Erosion was

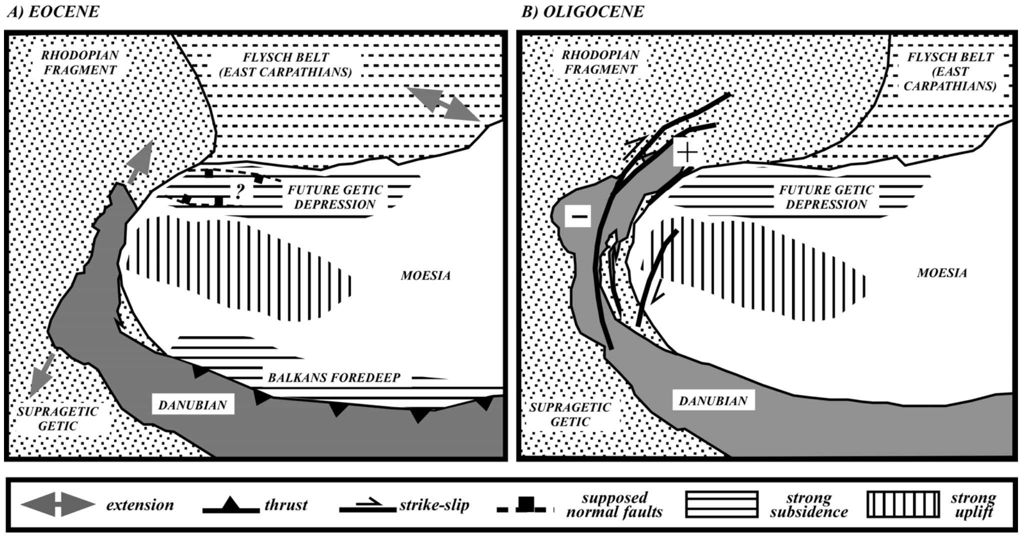


Figure 6.1 *Tectonic models for the Early Tertiary (Paleogene) evolution of the Carpathians region.* ***A)*** *Eocene;* ***B)*** *Oligocene. The tectonic framework is largely based on the model proposed by Fügenschuh and Schmid (submitted).*

the consequence of the uplift of a flexural foreland arch due to compressional intraplate stresses reflecting the strong mechanical coupling between the Balkans and Moesia. This uplift-driven erosion continued during Oligocene times (Fig. 6.1B).

Unloading of the western margin of Moesia due to core-complex formation could have contributed to the erosion by flexural rebound. Also, the dextral movement probably resulted in detachment of potential subducting slab of the South-Carpathians and corresponding unflexing of the foreland lithosphere (P. Ziegler, 2003, written communication). Elsewhere, the foreland (eastern Moesia and East-European/Scythian platform) remained during the entire Paleogene a stable region, with either continental or shallow marine environments.

6.3. EARLY MIOCENE (BURDIGALIAN)

The Burdigalian represents the last stage in the rotation of South-Carpathians around Moesian corner (Fig. 6.2A). Extension still took place in the South-Carpathians orogen (Maţenco and Schmid, 1999; Fügenschuh and Schmid, submitted), but also within the N-NW margin of Moesia where rift basins formed (Răbăgia and Maţenco, 1999; Maţenco et al., 2003). Extensional structures may be also present below the eastern part of the present Getic Depression, but it is difficult to be more precise because later contractional deformations and burial below younger sediments have obscured any evidence. For the region between the E-most part of Getic Depression and the external SE Carpathians (present SE Bend), the thick Burdigalian-Badenian sequence (1.8-2.3 km) suggests that an extensional event followed the Early Burdigalian salt deposition, according to Ştefănescu et al. (2000).

Extension in the South-Carpathians/N-NW Moesia is coeval with the S-wards thrusting of the internal East-Carpathians over the northern margin of the Transylvanian basin and the development of an adjacent retro-wedge foredeep basin (de Broucker et al., 1998; Ciulavu, 1999) in front of the Piennides thrust system (Săndulescu, 1884). Dextral strike-slip and oblique thrusting is suggested in the East-Carpathians nappes (e.g. Maţenco, 1997; Maţenco and Bertotti, 2000).

During Burdigalian times, large-scale erosional features evidence major uplift of western Moesia. This erosional stage seems to reflect the uplift of the southern “rift” shoulder of the Burdigalian extensional basins. Elsewhere, the Carpathians foreland remained as a low and rather stable, continental (non-depositional) area.

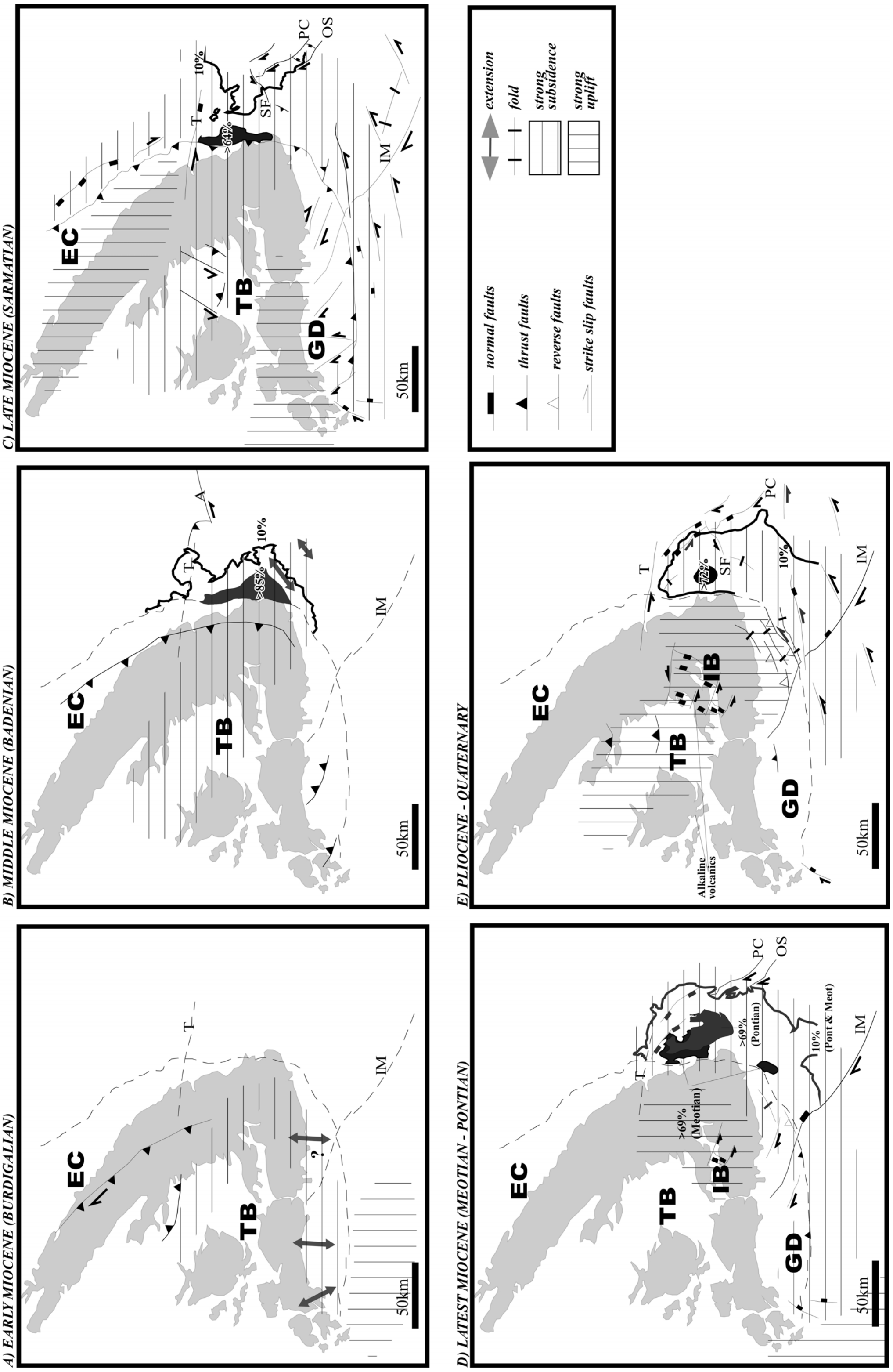
6.4. MIDDLE MIOCENE (BADENIAN)

During Badenian times, thrusting was active in the East-Carpathians and to the west of the present Getic Depression in response to regional NE-SW contraction (Fig. 6.2B; e.g. Maţenco and Bertotti, 2000; Maţenco et al., 1997b). At the same time, compressional stresses were exerted on the East-Carpathians foreland (East-European/Scythian platform) as evidenced by the activation of a system of thrust and wrench faults. This suggests mechanical coupling (Ziegler et al., 2002) between the strong East-European/Scythian platform and the encroaching orogenic wedge. However, there is no evidence for compression in Moesia; instead, NE-SW extension is documented for its eastern part. Extensional structures formed from at least the eastern margin of the Focşani Depression to outcropping Dobrogea (Hippolyte, 2002). A possible continuation of the Focşani extensional basin beneath the present Carpathians structures could be documented.

In terms of vertical movements, the Badenian time span was characterized by strong subsidence in the Focşani Depression and the Transylvanian basin (see for instance Ciulavu et al., 2000). A continuous NW-SE oriented subsiding area is inferred, which extended from the Transylvanian basin to the Focşani Depression (eastern Moesia). This broad and elongated subsiding area was flanked by the East- and South-Carpathians, uplift of which began towards the end of Badenian times (Sanders et al., 1999).

North of the Trotuş fault, the East-European/Scythian platform recorded only minor subsidence. In these areas, widespread, thin clastics mixed with evaporites suggest a rather stable region. West of the extensional basin, i.e. in western Moesia, the previous uplifting areas began to subside slowly, as evidenced by the onset of infilling of pre-existing erosional relief.

Figure 6.2 (next page) *Tectonic model for the Neogene evolution of the Carpathians region and subsidence patterns in the Focşani Depression:* ***A)*** *Early Miocene (Burdigalian);* ***B)*** *Middle Miocene (Badenian);* ***C)*** *Late Miocene (Sarmatian);* ***D)*** *Latest Miocene (Meotian-Pontian) and* ***E)*** *Pliocene-Quaternary. The dark gray spots represent areas with major subsidence in Focşani Depression. The percentages refer to the subsidence rates in Focşani Depression relative to their maximum value for each time frame. The present-day elevation contour of 500 m is shown in background for each time frame. Abbreviations of the foreland faults are: A Adjud fault; IM Intramoesian fault; OS Ostrov-Sinoe fault; PC Peceneaga-Camena fault; SF South-Focşani fault; T Trotuş fault. Other abbreviations: EC East Carpathians; GD Getic Depression; IB Intramountain basins; TB Transylvanian basin.*


A) EARLY MIOCENE (BURDIGALIAN)
B) MIDDLE MIOCENE (BADENIAN)
C) LATE MIOCENE (SARMATIAN)
D) LATEST MIOCENE (MEOTIAN - PONTIAN)
E) PLIOCENE - QUATERNARY
EC
TB
GD
IB
T
IM
PC
OS
SF
50km
10%
>85%
>64%
>69% (Meotian)
>69% (Pontian)
10% (Pont & Meot)
>72%
Alkaline volcanics
normal faults
thrust faults
reverse faults
strike slip faults
extension
fold
strong subsidence
strong uplift

6.5. LATE MIOCENE (SARMATIAN)

During the Sarmatian, major contractional deformation occurred in the East-Carpathians where the Subcarpathian nappe was finally thrust onto the foreland (Fig. 6.2C; e.g. Săndulescu, 1984, 1988). Shortening ended in the Middle Sarmatian north of the Trotuş fault and in the Late Sarmatian between the Trotuş and Intramoesian faults (e.g. Săndulescu, 1984; Maţenco and Bertotti, 2000). At the same time, mostly dextral strike-slip faults deformed the Getic Depression, which was thrust onto the northern Moesian margin to the west of the Intramoesian fault (Maţenco et al., 1997b; Răbăgia and Maţenco, 1999). In the hinterland (Transylvanian basin), sinistral strike-slip faulting and transpresional structures developed (Ciulavu et al., 2000).

Along strike, the Sarmatian deformation pattern of the foreland varies considerably, despite of the essentially coeval emplacement of the Subcarpathian nappe. Orogen-parallel normal faulting with decreasing offsets from N to S affected the East-European platform. Southwards, limited dextral shearing may have occurred in the foreland (A. Răbăgia 2001, personal communication). The N to S decrease of those normal fault offsets correlates with a N to S decrease of the magnitude of shortening in the orogenic wedge, and with the hard and soft collisions that took place in the north and in the south, respectively (Maţenco, 1997). At the same time, the entire East-European/Scythian platform was tilted towards the East-Carpathians, thus forming a typical foredeep.

By contrast, the amount of tilting of eastern Moesia was relatively minor, except for the region strictly confined to the Focşani Depression. Along its eastern and southeastern flanks, Sarmatian deformation was dominated by strike-slip faults (Chapter 4, fig. 4.11). Separated by the Intramoesian fault, the western Moesian block was tilted N-wards similar to the East-European platform. Although some minor orogen-parallel normal faulting could be related to the emplacement of the Subcarpathian nappe, the largest fault systems formed in the westernmost part of Moesia in response to transtension (Răbăgia and Maţenco, 1999) rather than to contraction.

The Sarmatian vertical movement pattern is characterized by a NW-SE oriented zone of subsidence, which extends from the Transylvanian basin to the Focşani Depression (Fig. 6.2C). However, as in the Carpathians Bend zone shortening and thrusting partly compensated basin floor subsidence, the connection between the Transylvanian and Focşani marine basins was more reduced than during Badenian times. The subsiding zone was flanked to the NE and SW by bands of very strong exhumation and uplift. Fission track data indicate that the exhumation of the East-Carpathians to the north of the Trotuş fault and of the South-Carpathians commenced towards the end of the Badenian and continued during Sarmatian times (Sanders et al., 1999). Sediment supply directions documented in the foreland support these exhumation ages.

6.6. LATEST MIOCENE (MEOTIAN-PONTIAN)

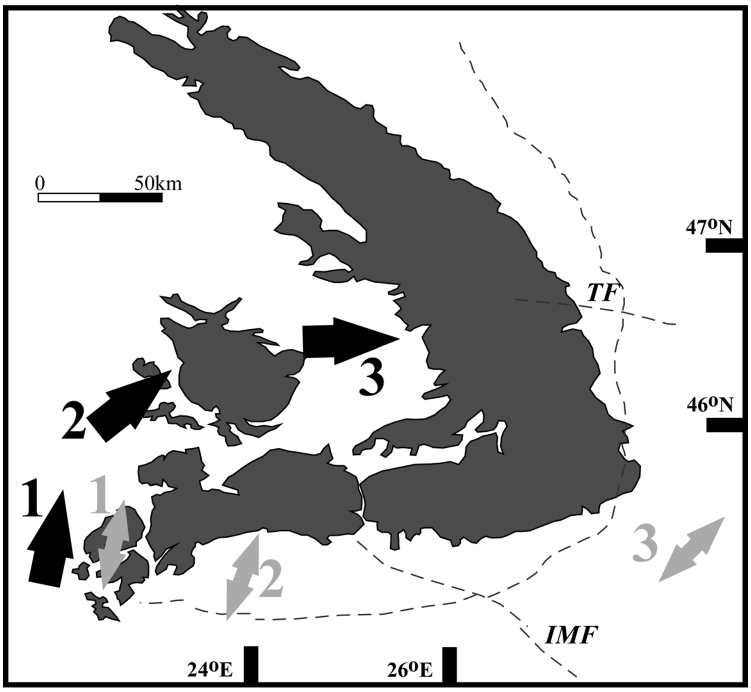
By this time, the Carpathians belt was already structured and no major shortening occurred during the Meotian-Pontian time span (Fig. 6.2D). However, within the Bend area, minor NW-SE dextral strike-slip and thrusting occurred during the Meotian (paleostress data in Hippolyte and Săndulescu, 1996; Maţenco and Bertotti, 2000). This was accompanied by an ESE advance of this orogenic wedge, which commenced during the Latest Sarmatian, that is, just after the emplacement of the Subcarpathian nappe (Maţenco and Bertotti, 2000). During the Pontian, these movements of the Bend zone orogenic wedge diminished or ceased.

During the Meotian, in the eastern part of the Getic Depression minor reverse and dextral strike-slip faults were active (Maţenco et al., 1997b; Răbăgia and Maţenco, 1999), which can be correlated with movements along the Intramoesian fault (Tărăpoancă, 1996). To the internal part of the Carpathians Bend zone, opening of tensional intramountain basins began in the Pontian (Ciulavu, 1999) at the rear of the ESE-wards moving orogenic wedge. From the Sarmatian onwards, no apparent deformation can be identified on the East-European/Scythian platform. In the eastern part of Moesia, minor reactivation of the Intramoesian and Peceneaga-Camena faults is evident. The eastern margin of the Focşani Depression experienced normal faulting. During the Meotian-Pontian, the Focşani Depression continued to subside rapidly, albeit at somewhat lower rates than during the Sarmatian. Similarly, the NW Moesia subsided rapidely (Maţenco et al., 2003; Bertotti et al., 2003). Sedimentation ceased on the East-European/Scythian platform whereas the Transylvanian basin slowly subsided only during the Meotian (Ciulavu et al., 2000). Uplift and exhumation of the Carpathians Bend zone began towards the end of the Pontian (Sanders et al., 1999), separating the Transylvanian from the Focşani basin, which previously had formed part of the same subsiding area. Apparently, uplift of the East- and most of the South-Carpathians was not associated with major deformation (Sanders et al., 1999). However, in the Pontian, large deltas in the western part of Moesia prograded from the W, SW and S suggesting that the South-Carpathians orogen either was uplifted and/or that by this time, the Danube had broken through it. If the former suggestion applies, it appears that the onset of exhumation of both curved sectors of the Carpathians belt occurred simultaneously.

6.7. PLIOCENE–QUATERNARY

Contractional deformation resumed in the Bend area after Pontian times (Fig. 6.2E; e.g. Hippolyte and Săndulescu, 1996). Folding and faulting occurred coeval with the steepening of the western margin of the Focşani Depression. In the internal part of the Bend zone, tensional opening of intramountain basins continued along WNW-ESE sinistral strike-slip faults (Ciulavu, 1999). Minor contractional deformation is documented in the eastern part of Getic Depression (Maţenco et al., 1997b) and within the Transylvanian basin (Ciulavu et al., 2000). Uniform tilting and transtensional reactivation of the Intramoesian and Peceneaga-Camena faults accompanied continuing subsidence of the Focşani Depression and the eastern Moesia. Very young ENE-WSW trending faults in large areas of Moesia may be interpreted as sinistral strike-slips (also Răbăgia and Tărăpoancă, 1999; Răbăgia et al., 2000). Normal/transtensional faulting is documented in the foreland of the curved sectors of the Carpathians belt, namely along the eastern margin of the Focşani Depression and in the westernmost part of Moesia (Fig. 6.2E). During this time span, the area of maximum subsidence gradually became almost circular, occupying the central part of Focşani Depression. Here and south of the Focşani basin, a large increase in the subsidence rate is recorded. Subsidence of the Focşani Depression was coeval with the exhumation of the entire SE Carpathians Bend zone (Sanders et al., 1999). Generally, foreland subsidence was confined only to Moesia. Simultaneously, the Transylvanian basin became uplifted (Ciulavu et al., 2000).

Figure 6.3 (next page) *Changes in the tectonic transport directions of the Carpathians units during: 1) Eocene-Oligocene (after Fügenschuh and Schmid, submitted); 2) Early-to-Middle Miocene and 3) Middle-to-Late Miocene (after Hippolyte et al., 1999) and migration of extension from the orogen to the lower plate. The present-day elevation contour of 500 m is shown for reference. IMF Intramoesian fault, TF Trotuş fault.*



6.8. FORELAND EXTENSION IN A REGIME OF OVERALL CONVERGENCE

Since the Paleogene, the Rhodopian fragment (sensu Burchfiel, 1976) rotated around the western Moesian corner and moved towards the NE-E. Tectonic transport directions changed through time showing a clockwise rotation since the Eocene until the Late Miocene (Fig. 6.3; e.g. Ratschbacher et al., 1993; Schmid et al., 1998). During this time span, it appears that extension affected regions progressively farther to the east. The extension direction appears to follow the clockwise rotation, from the South-Carpathians to the eastern Moesia and to migrate from the inner part of the orogen to the foreland. A genetic link between convergence direction of the upper plate and the site of extension in the foreland plate is suggested. In other words, oblique convergence between the upper plate and Moesia during the Paleogene–Middle Miocene (Badenian) may have promoted extension on the northern margin of the latter.

By contrast, frontal collision took place between the upper plate and the East-European/Scythian platform. Syn-collisional compressional deformation of the foreland indicates a mechanical coupling between the upper and the foreland plates. The rheological properties of the two lithospheric blocks of the Carpathians foreland plate (East-European/Scythian platform and Moesia) seem to control the magnitude of strain during convergence and the amount of post-shortening foredeep subsidence.

The process of clockwise rotation and dextral translation of the Carpathians units around the Moesian corner, as well as the deformation pattern recorded by the foredeep could get additional insights if they are compared with the Caribbean plate-tectonic setting, where the E-wards displacement of the Caribbean plate relative to the South-America seems to present some similarities with the kinematics of the South-Carpathians/Moesia boundary (H. Doust, 2004, personal communication). The geometry of the present-day Caribbean plate and the plate-tectonic framework resembles fairly well that of the Carpathians (Fig. 6.4A). The Caribbean plate has moved from the Late Cretaceous onwards towards east, overriding in its

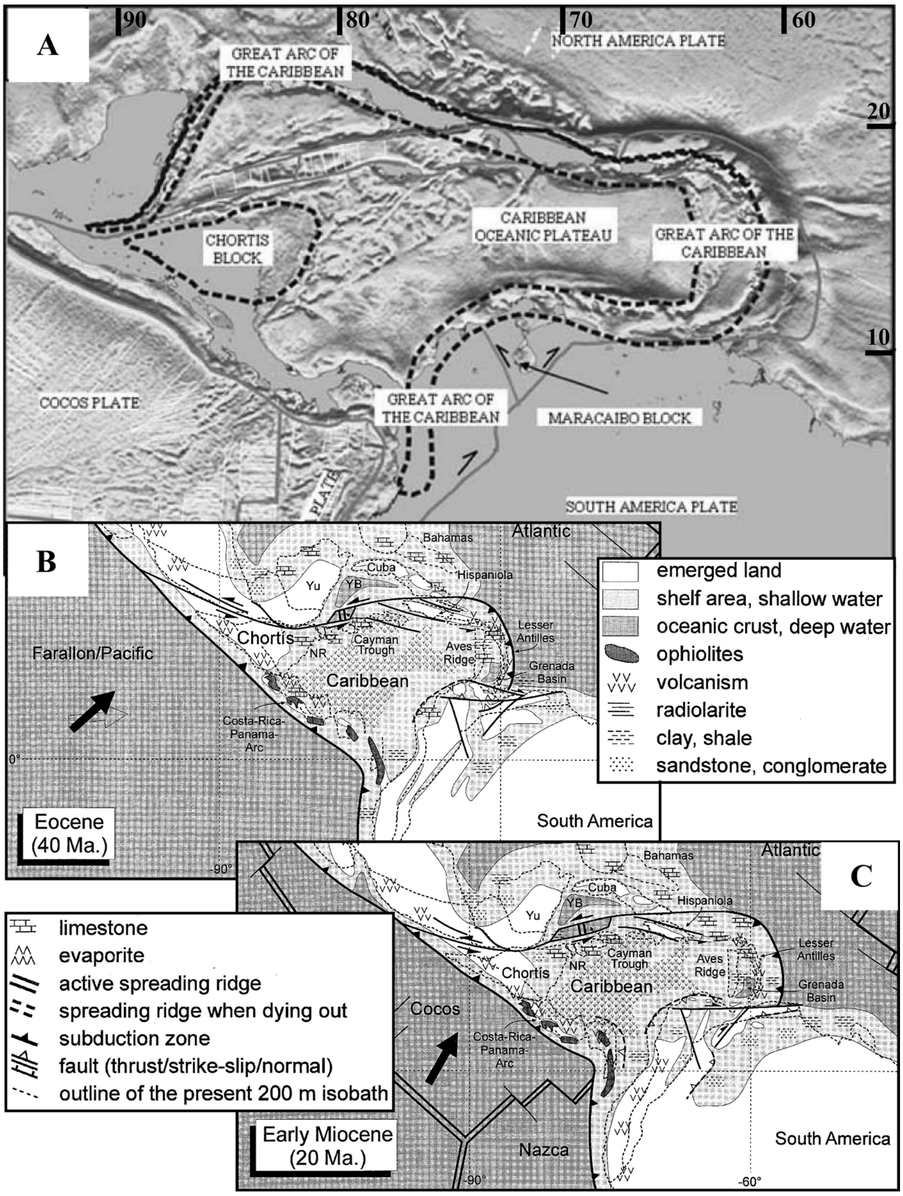


Figure 6.4 *(**A**) Plate-tectonic setting of the Caribbean region. (**B**) and (**C**) show the Eocene and Early Miocene, respectively, restored positions of the Caribbean and surrounding plates and the major deformations produced at the Caribbean boundaries (Meschede and Frisch, 1998).*

frontal part the Atlantic plate along the Lesser Antilles arc (e.g. Figs. 6.4B and C), whereas its movement relative to the North- and South-Americas continental plates, respectively, has been accommodated by sinistral and dextral transform zones (e.g. Pindell, 1991; Meschende and Frisch, 1998). Succesive pull-apart basins, subsequently inverted, and foredeep basins younging towards east were formed along the northern South-America margin due to the progressive displacement of the Caribbean plate that induced S-vergent thrusting (Pindell, 1991 and references therein). This indicates a dominant transpressional boundary, which is similar to the tectonic regime that characterized the Getic Depression during Badenian and Sarmatian times (also see Răbăgia and Mațenco, 1999). However, in the South-Carpathians/Moesian boundary setting, there is no clear evidence for E-wards younging of the foredeep basin. Otherwise, such a trend is observed in the locus of foreland extension, which may indicate a dominant transtensional mechanism for the overall normal faulting along the northern margin of Moesia.

When compared with the Caribbean/South-America plate-boundary deformations, an apparent weakness of the model of the South-Carpathians kinematics arises, which refers to the absence of any contraction in the NW corner of Moesia, that, instead, recorded normal faulting and great subsidence during the E-wards movement of the orogenic units. Also, this model assumes basically no translations of western Moesia towards the South-Carpathians, although the Intramoesian fault is demonstrated to have been dextrally active.

It will be worth to test through either numerical or analogue modeling, the pattern of deformations that would result from coupling the E-wards translations of the South-Carpathians with the NW-wards displacement of western Moesia, especially for the Sarmatian times.

In the Caribbean plate-tectonic setting, the strong compression exerted on the NW corner of South America determined the N-wards extrusion of the triangular Maracaibo block (Figs. 6.4A-C), which created the South Caribbean fold-and-thrust belt (Pindell, 1991 and references therein). Although no such deformations have been identified so far in the South-Carpathians/Moesian corner, this might have happened in Paleogene times, as such a NE-SW dextral strike-slip fault seems to have been active then in westernmost part of Moesia (e.g. Fig. 3.2).

6.9. SUBSIDENCE MECHANISMS

The two stages of extension that affected the northern part of Moesia are of primary importance for the onset of sediment accumulation in the Carpathians foredeep, as well as for its subsequent subsidence history. In western Moesia, the Burdigalian subsidence is clearly related to crustal extension (Răbăgia and Mațenco, 1999). Although the magnitude of Burdigalian (and Paleogene?) extension is not yet constrained, it is likely that post-rifting subsidence played in this area a significant role. As western Moesia was characterized prior to and during rifting by relatively high elevations, the post-rift subsidence is probably responsible for the onset of regional sedimentation during the Badenian.

Badenian extension affecting eastern Moesia has been addressed quantitatively in Chapter 5, indicating that Badenian subsidence can be fully explained by small amounts of extension. In contrast with the extension from western Moesia, no significant thermal cooling subsidence is predicted for its eastern part.

Part of the post-extensional subsidence is genetically related to foreland flexure, triggered by the emplacement of the Subcarpathian nappe during the Sarmatian. Orogenic loading computed from the present topography explains relatively well the foredeep basin overlying East-European/Scythian platform, as well as the depth of the Transylvanian basin. Localization of the major subsiding areas is mainly the effect of the 3D distribution of orogenic load and lateral changes in the flexural strength of the lithosphere.

Significant post-shortening subsidence is recorded only in Moesia, particularly in the Focșani Depression. Neither young strike-slip nor normal faulting can totally account for the magnitude of these vertical movements. An extra-topography of 500 m height adopted in the SE Carpathians Bend zone in the planform flexural model, together with the present topography, can explain most of the subsidence of the Focşani Depression and the wavelength of lithospheric deflection. An oceanic or delaminated continental mantle slab attached at the base of the lithosphere beneath Vrancea zone (models cited in subchapter 2.4) would significantly overestimate both the depth and wavelength of the observed Focşani basin. Moreover, the presence of such a slab is not consistent with the location of depocentral areas of the Carpathians foredeep basin and with the pattern of vertical movements recorded in this orogen/foreland system (also see Maţenco et al., 2003; Bertotti et al., 2003). Rather, the required extra-topography could represent a thrust load neglected due to the use of "static" (instantaneous loading) instead of planform flexural kinematic modeling.

In addition to extension and flexure, lithospheric buckling is proposed (see also Bertotti et al., 2003) as a third mechanism to explain the remaining subsidence (maximum of 2 km in the Focşani Depression). The onset of buckling would have been Pliocene (end Pontian?) and is suggested by the pattern of vertical movements (Fig. 6.2E).

6.10. CHANGES IN RELATIVE SEA LEVEL RECORDED IN THE CLINOFORM PATTERNS

During and following the emplacement of the outermost Carpathians nappe onto the foreland, large deltas transporting the sediments from the orogen to the foreland basin developed in various places and at different times. Creation of the accommodation space can be related to different tectonic mechanisms active throughout the Late Miocene-Pliocene and led sometimes to significant deepening of the basin floor. Based on the seismic facies of the prograding sequences (presented in Chapters 3 and 4), the spatial and temporal distribution of these deltas, as well as the relative depth of the receiving basin is shown schematically in Figure 6.5.

First deltaic system prograded SSE-wards along the central-northern East-Carpathians foredep axis during the Sarmatian (also Negulescu, 2001), into a relatively deep water environment (~400-500 m), which occupied, at least, the northern domain of the Focşani Depression. The deltaic system was formed in response to thrusting and accelerating exhumation of the central-northern East Carpathian orogenic wedge.

North-south transport of sediments from the Carpathians to the foreland basin continued during the Meotian. Prograding sequences of this age can be observed on a much larger area than during Sarmatian times, and occupied the northern part of most of the South-Carpathians foreland basin. This time, however, the dip of the clinoforms indicates everywhere a shallow water environment. In contrast to the Sarmatian progradation, the Meotian transport direction was roughly transversal to the foredeep axis. In the northern foreland (East-European/Scythian platform), the previous thrusting-induced rise of the relative sea level was followed by a continuous drop due to the progressively filling of the basin in the absence of any significant tectonic subsidence.

The Pontian represent a moment when another significant rise in relative sea level is recorded, this time in the NW-most part of Moesia. This basin was fed by deltaic systems transporting sediments from the west and the east. In contrast to the Sarmatian thrusting-related rise, the deepening of the basin floor in NW-most Moesia was a consequence of normal faulting.

The last stage in development large-scale deltaic systems occurred in Pliocene times, when the dominant direction of the sediment supply was WNW-ENE (also Jipa, 1997). The sediments were supplied mostly from the South-Carpathians, but also from the SE Carpathians orogenic wedges, to a Moesian shallow water lake. This ~E-wards large-scale progradation has continued until present, when the accumulation space for the sediments transported from

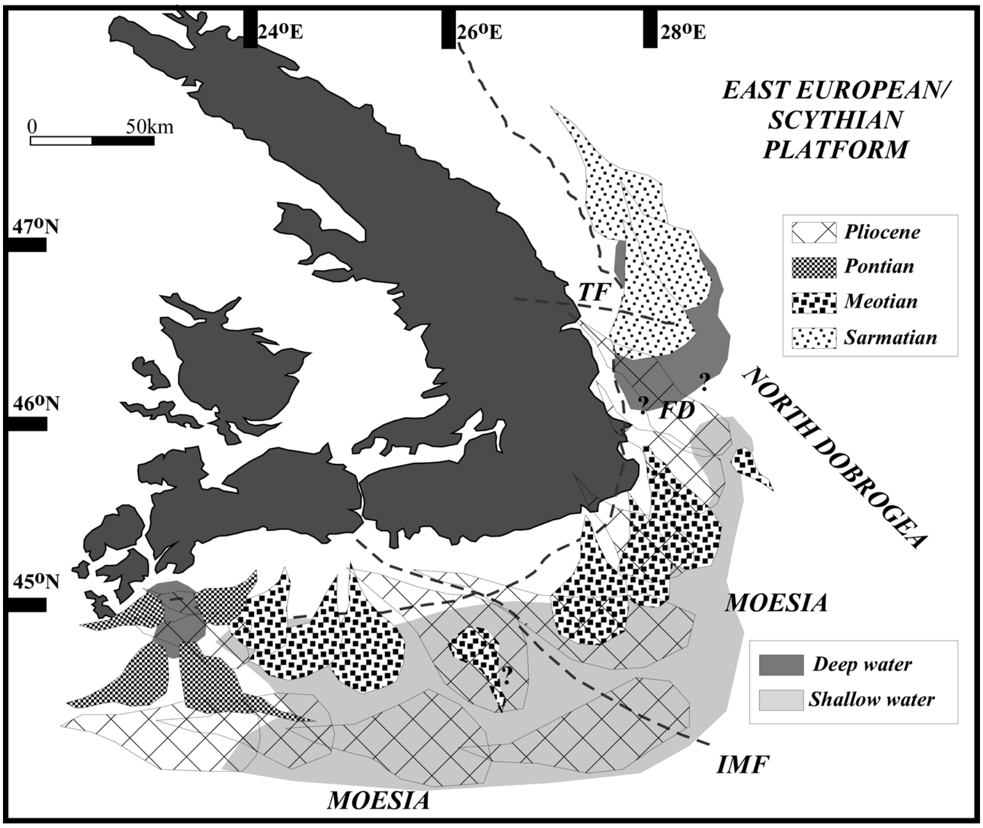


Figure 6.5 *Major prograding systems in the Carpathians foredeep and the relative bathymetry of the sediment accumulation basin, inferred from the clinoforms architecture (discussion and additional references in text). The present-day elevation contour of 500 m is shown for reference. FD Focşani Depression; IMF Intramoesian fault, TF Trotuş fault.*

the Carpathians is provided by the Black Sea, and the present drainage network roughly coincides with the Pliocene one.

Migration of the sedimentary prograding systems along the foredeep, from the NE to the SW, from the Sarmatian to the Pontian, and subsequently E-wards from the Pliocene onwards, appears to reflect the complex interplay between the dynamics of the source areas in terms of tectonic topography (Cloetingh et al., 2003) and the creation of accommodation space that essentially depends on the “rheological response” of the foreland plate to the tectonic processes, such as flexural orogenic loading and lithospheric buckling.

6.11. GENERAL CONCLUSIONS AND FINAL REMARKS

The Carpathians orogen represents one of the most arcuate segments of the Alpine belt that acquired its shape mostly during the Tertiary. A key role in the tectonic history of the Romanian Carpathians was played by the behaviour of the foreland plate, which is made up of two major lithospheric blocks that differ in rheological properties and crustal/sub-crustal mantle thicknesses, namely the East-European/Scythian platform and Moesia. From the East-Carpathians frontal thrust to the SE, an intervening lithospheric block is formed by the North Dobrogea orogen that behaved rheologicaly quite similar as the East-European platform during Tertiary times. Major crustal faults bound these different foreland blocks and project beneath the Focşani Depression that is located in front of the SE Carpathians Bend and forms the most subsiding part of the entire Carpathians foredeep.

In terms of tectonic deformation and vertical movements, the foreland plate responded spatially differently to the encroaching Carpathians orogenic wedge. The Tertiary deformation and subsidence patterns recorded in the East- and South-Carpathians foreland basin show major, sometimes even contrasting differences between the East-European/Scythian platform and Moesia.

The East-European/Scythian platform behaved as a typical foredeep/foreland and subsided flexurally during the Middle-to-Late Miocene thrust emplacement. After the final Middle Sarmatian thrusting, both the orogenic wedge and the foreland were uplifted and eroded. On the other hand, the evolution of the Moesian sector of the foredeep is characterized by changing patterns of structural deformation, uplift and subsidence, including tilting towards the orogen. The northern margin of Moesia was affected by extension from the South-Carpathians to the southern edge of the East-Carpathians. Anomalous Tertiary patterns of vertical movements are recorded in Moesia, from over 2 km uplift in its western part to ~13 km subsidence in the Focşani Depression. From the Badenian onwards, the site of major subsidence characterized the Focşani Depression. In contrast to the East-European/Scythian platform, where subsidence ceased towards the end of the Sarmatian, Moesia continued to subside. After the main shortening event, which was coeval in the East- and South-Carpathians (Middle-Late Sarmatian), additional shortening occurred mainly in the SE Carpathians Bend zone and was accompanied by faulting in the Moesian part of the foreland.

Tectonic activity in the foreland of the Carpathians Bend, involving wrench and normal faulting commencing during the Sarmatian and persisting until the present, indicates strong mechanical coupling of the orogenic wedge with a pre-fractured foreland.

Overall, the strongest subsiding areas of the entire Carpathians foredeep/foreland trend are located in front of the concave and convex sides of the Carpathians, i.e. NW Moesia and Focşani Depression. Although so far, no direct connection has been established between the subsidence of the foredeep and accompanying tectonic deformation, it is noteworthy that the internal parts of the Carpathians bends represent also the sites of the youngest (Quaternary) alkaline volcanism (Fig. 6.2E) probably sourced from the mantle.

The differences in tectonic behavior between the two major lithospheric blocks forming the Carpathians foreland plate are primary related to their rheological properties, i.e. strong East-European/Scythian platform versus weak Moesia, and to the convergence direction of the Carpathians with the foreland, i.e. frontal for the East-European/Scythian platform and oblique for most of Moesia.

Quantitative modeling of the subsidence in the Carpathians foredeep, particularly in the Focşani Depression, shows that its shape, position and depth could be explained by the succession of three different mechanisms, namely extension, flexure and lithospheric buckling. In contrast to previous works linking the subsidence and the last stages of the

Carpathians evolution exclusively to deep-seated effects of a lithospheric slab, this Thesis highlights the importance of (1) separating different subsidence stages, (2) lateral variations in lithospheric strength and (3) the 3D effect of the topographic load.

Although most of the present tectonic activity occurs in the SE Carpathians Bend (Vrancea zone) with earthquakes at depths ranging down to 200 km, shallower earthquakes, mainly crustal, have been recorded over large parts of Moesia (e.g. Cornea and Lăzărescu, 1980). The pattern of the ground horizontal acceleration recorded for the last large (magnitudes >6) earthquakes (Mândrescu, 1995) shows that the highest values have either a NE-SW-trend in the central-eastern Moesia or a NW-SE-trend in NE Moesia/North Dobrogea promontory, correlating well with the ENE and NW-trending faults, respectively. Indeed, the reflection seismic data indicate that these fault systems are still active. These findings should be seen as a base for new models explaining/forecasting the seismicity in this part of the Carpathians region, as well as for urban planning.

Documenting the pattern of vertical movements, as well as the subsidence mechanisms of the Carpathians foredeep could be of major importance to hydrocarbon exploitation/exploration studies. Although the Romanian petroleum industry looks back on a history of more than 100 years, studies focused on basin analysis are scarce. In any approach dealing with the petroleum potential of a sedimentary basin, one of the most sensitive topics and, at the same time, one of the most difficult to constrain, is the thermal history, as it controls the maturation level of source rocks. The inferred ages of genesis and expulsion of different hydrocarbon phases are dependent on vertical movements-induced thermal changes. Taking just the present heat flow (basically well known) and building a burial history of sediments derived only from their actual thicknesses may lead in some cases to wrong

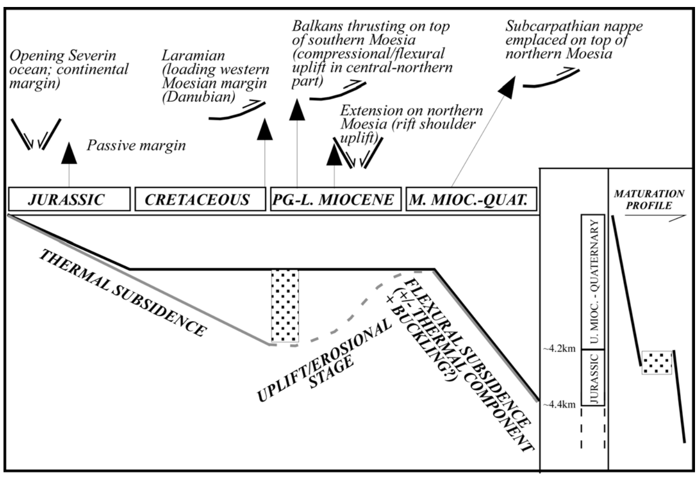


Figure 6.6 *Burial history plot (dark line) from a well drilled in the W-NW part of Moesia. Also shown schematically are the sedimentary column and the profiles of the organic metamorphism (from Paraschiv and Balteş, 1983). An important break in the maturation of organic matter is recorded (stippled rectangle). This break is probably related to a deeper burial than actually observed due to the sedimentary column deposited during the Cretaceous, which was subsequently removed by erosion (gray line). When deposition restarted in the Upper Miocene, the sediments underneath the unconformity were already experiencing heating. Consequently, their maturity is higher than that of the young sediments.*

conclusions, and hence, to the adoption of an inappropriate exploration strategy. For instance, in W-NW Moesia, the long-lasting erosional period, which removed ~2 km from the sedimentary column, should have had a significant impact on the thermal history of potential source rocks. Indeed, a maturation analysis undertaken on samples in wells drilled in this region shows major breaks in the organic metamorphism profiles beneath the Neogene sequence (Paraschiv and Balteş, 1983). This pattern is mostly related to the burial of the sedimentary column at depths of ~2 km-deeper than their present-day position prior to the onset of erosion (Fig. 6.6). Approximately in the same region, a modeling study focused on the maturity of potential source rocks (Pene, 1996), which did not take into account the erosional period and used only the present heat flow, concluded that the Paleozoic and Mesozoic organic matter-bearing sediments are mostly immature.

Since the youngest tectonic deformations of the Romanian Carpathians foreland occur basically in Moesia, a correlation with coeval structures in the regions located farther south is necessarily. As described by Picha (2002) for the Dinarides foredeep and the Hellenides, orogen-parallel strike-slip faulting characterizes the post-collisional tectonics in this part of the Alpine realm. Picha (2002) interpreted this young (Pliocene-Quaternary) strike-slip faulting as the response of the orogen/foreland system to regional compressional stresses after the foreland thrust propagation became locked. A similar explanation may apply for Moesia because most of the deformation documented here is strike-slip and the fault systems generally parallel the orogen.

Because this Thesis is focused mainly on tectonics and subsidence modeling, a study dealing with the sequence stratigraphy of the Carpathians foreland basin appears as the next and, at the same time, very useful step. It could add more information to compute accurately the tectonic subsidence of the basin, as a precise chronology of the relative sea level variations is lacking. Spatial and temporal changes in the sedimentation paleo-environments, as well as precise lithological and petrophysical mapping will provide good input data necessary to construct decompaction curves and water depth profiles.

Further modeling studies are proposed, especially for the mechanism of the major Eocene-Burdigalian uplift of western Moesia and the synchronous significant subsidence in the South-Carpathians foredeep. In this Thesis, the proposed explanation for the major uplift-driven erosion has been related to the succession of two relatively low-scale mechanisms occurring in the same region, one contraction-related and the other extension-related. Taking into account the strongly arcuate shape of the plate boundary, a planform flexural kinematic study coupled with an extensional modeling would be suitable to test this hypothesis. The reversal of vertical movements, which started during Badenian times, indicates that the amount of subsidence was higher than that derived from the sediment thicknesses. In the South-Carpathians foredeep, in contrast with eastern Moesia, thermal subsidence following the Burdigalian rifting may have played a significant role. However, the first step should be to document the Paleogene setting (deeply buried beneath Neogene deposits) because the large thickness of sediments (~5 km, Jipa, 1980, 1984) may be also related to an extensional stage. If this were true, the thermal relaxation prior to and during the flexural stage could possibly account for the subsidence of the South-Carpathians foredeep, as demonstrated for other foreland basins (e.g. Deségaulx et al., 1991).

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